Appalachian State University
Boone, North Carolina

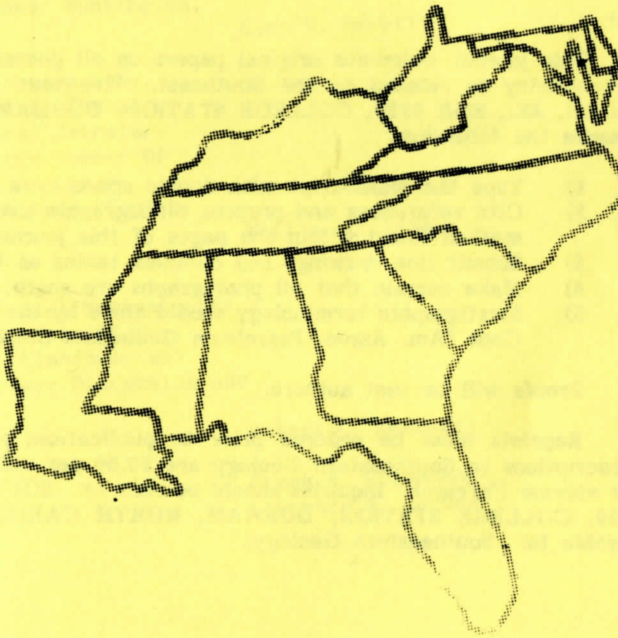
Southeastern Geology: Volume 26, No. 3
February 1986

Edited by: S. Duncan Heron, Jr.

Abstract

Academic journal published quarterly by the Department of Geology, Duke University.

SOUTHEASTERN GEOLOGY



PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 26, NO. 3 FEBRUARY, 1986

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

Editor in Chief:
S. Duncan Heron, Jr.

Managing Editor:
James W. Clarke

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SOUTHEASTERN GEOLOGY

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A MESOSCOPIC THRUST SYSTEM IN WEST VIRGINIA: ITS DEFORMATION HISTORY AND REGIONAL IMPORTANCE

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ABSTRACT

First-order parasitic folds in the sub-vertical western limb of Wills Mountain anticline deform an older mesoscopic thrust system. The system consists of small ramps connected by extensive flats, is greater in extent than the folds, and causes greater shortening than the folds, unlike other pre-folding contraction faults. Structural geometry is locally complicated by a folded ramp in an anticlinal core and syn-folding interlayer-slip in only the footwall. The system is probably older than the host Wills Mountain anticline, having formed during earlier piggyback propagation from the Cambro-Ordovician level in the Appalachian Foreland thrust belt.

INTRODUCTION

A major problem in the central Appalachian Foreland thrust belt is a possible shortening difference between lithotectonic units or levels. On one hand, shortening is accommodated at a lower level of the thrust system in the Cambro-Ordovician carbonates by large-scale imbrication (Gwinn, 1964; Perry, 1978a). However, at higher levels such large-scale imbrication is mostly absent. Also, folds above the lower-level imbrication produce insufficient shortening. Yet, if the system deformed uniformly, the total strain must be equal between the lower and higher levels.

Recently, structures other than large thrusts and folds have been shown to accomodate the necessary additional strain at the higher levels (Bowen, 1985; Geiser, 1985; Herman and Geiser, 1985). The shortening is mostly by smaller folds, faults, cleavage, and flattened grains (Herman and Geiser, 1985). The smaller, or mesoscopic, faults are considered to be the least important structures because they are frequently subsidiary to folds (Berger and others, 1979) and larger parent faults are uncommon. However, these mesoscopic fault systems provide very useful keys to unraveling local deformation history, and they may in fact have a major role in accommodating strain. This paper is concerned with the role and development of such mesoscopic fault systems, and will:

1) Describe a mesoscopic thrust system from a higher lithotectonic unit.
2) Demonstrate that the system is not subsiderary to folding, and
3) Describe the deformation of the system during subsequent folding and relate this to the regional deformation sequence.

REGIONAL GEOLOGY

The mesoscopic thrust system examined here is located in a roadcut along State Route 28 at the Grant-Pendleton County line, West Virginia (Figure 1). The locality is along the south side of a water gap, which was cut by the adjacent North Fork of the South Branch Potomac River.

The thrust system is in the western limb of the Wills Mountain anticline, the area's largest regional structure. This fold is the surficial expression of the westernmost completely duplicated thrust-horse of Cambro-Ordovician carbonates (Gwinn, 1964; Perry, 1978a). Also, the mesoscopic thrust system is deformed by first-order parasitic folds of the Wills Mountain anticline, including the Hopeville anticline and syncline. The folds contain Silurian and Lower Devoian rocks in a structural terrace. The terrace is decoupled from the underlying horse of Cambro-Ordovician carbonates across the weaker Ordovician Martinsburg or Reedswille Formation

(Perry, 1978b).

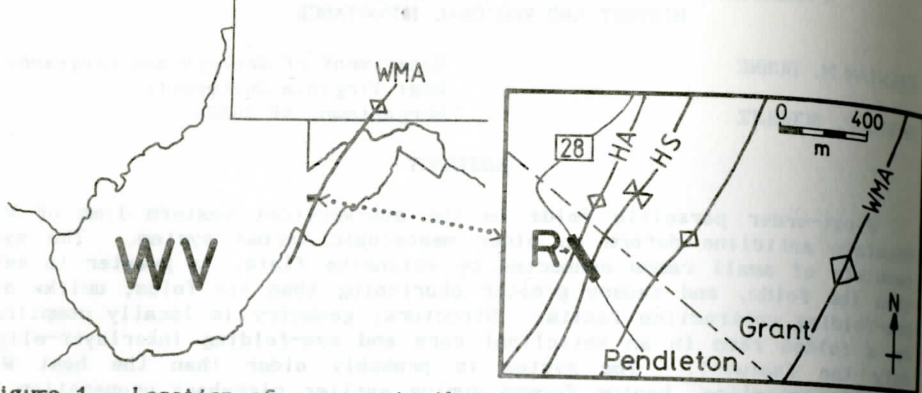

Figure 1. Location of mesoscopic thrust system (R) along State route 28, in West Virginia. Folds are WMA - Wills Mountain anticline, HA - Hopeville anticline, and HS - Hopeville syncline.

LOCAL STRATIGRAPHY

The thrust system displaces rocks of the Upper Silurian-Lower Devonian Helderberg Group (Head, 1969; 1974). The deformed sequence in the roadcut extends from the older Jersey Shore Limestone Member to the younger Corriganville Limestone (Figure 2). The lowest part of the Jersey Shore Limestone Member consists of interbedded terrigenous shales, shaly limestones, and some coral-stromatoporoid calcarenites (symbol 'm', Figures 2 and 3). The overlying part of the Jersey Shore Limestone Member consists of the distinctive, thicker-bedded, coral-stromatoporoid calcarenites grading

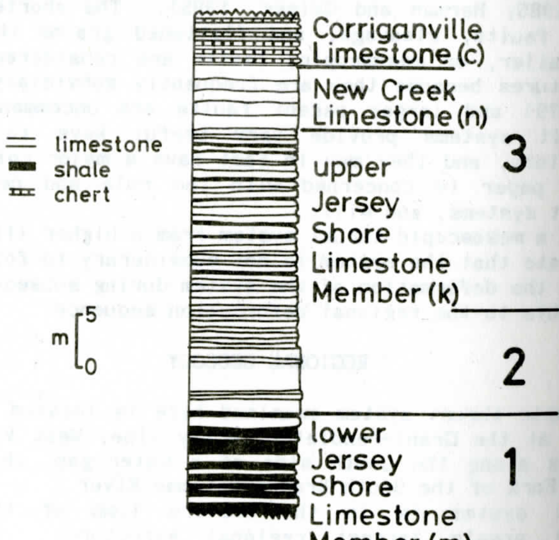

Figure 2. Diagrammatic lithologic column for lithostratigraphic and lithostructural units. Lithostratigraphy follows Head (1969; 1974).

upwards into the thinner-bedded, micritic calcilutites and calcisiltites (symbol "k", figures 2 and 3). Bioclastic calcarenites of the New Creek Limestone (symbol "n", Figures 2 and 3), and interbedded white cherts and limestones of the overlying Corriganville Limestone are at the top of the roadcut, above the thrust system (symbol "c", Figures 2 and 3).

This lithostratigraphy may be divided into lithological sequences that contain distinctive suites of small structures. Each suite represents a differing mechanical behavior during deformation, and consequently each lithological sequence constitutes a lithostructural unit (sensu Currie and others, 1962) (Figure 2).

The first lithostructural unit consists of the basal, interbedded terrigenous shales and limestones of the Jersey Shore Limestone Member. It contains disharmonic folds, shales with no cleavage but thickened by flowage, and bedding-parallel thrusts (Unit 1, Figure 2). The second unit consists of the thicker-bedded, coral-stromatoporoid calcarenites of the Jersey Shore Limestone Member. The unit has a massive appearance with a few contraction faults and locally abundant, bedding-parallel, calcite slickenside veins (Unit 2, Figure 2). The third unit consists of the remaining younger sequence. It contains kink folds verging toward the synclines, rootless anticlines above small contraction faults, small congruent parasitic folds, closely spaced contraction faults, and a few bedding parallel slip-surfaces (Unit 3, Figure 2).

Using the structural suites as strength indicators during deformation, Unit 1 was the weakest unit, with its flowage and disharmonic folding. Unit 2 was the strongest unit, with its stiffer limestones lacking folds or flowage. Unit 3 was of intermediate strength with its small folds and faults.

LOCAL STRUCTURAL GEOMETRY

The road cut contains a syncline-anticline fold pair, the Hopeville syncline and anticline (Figure 3). The Hopeville syncline (A, Figure 3) is a comparatively simple fold, which is upright-subhorizontal, rounded, gentle, and Class IB (Fleuty, 1964; Ramsay 1967). The axial plunge is 039/05. This syncline folds the thrust system, demonstrating that at least locally, the thrust system is older.

In the syncline, the thrust system consists of a network of flats and ramps that have less than 2 m cumulative displacement. The ramps are gentle with respect to bedding, since bedding-to-fault angles are less than 15 degrees. This staircase trajectory for the thrust system with ramps connected by extensive flats, occurs throughout the roadcut. The system does not consist of isolated imbricate thrusts, such as those for secondary pre- and syn-folding contraction faults (Berger and others, 1979). Also, all the ramps cut upsection to the northwest, independent of thrust dip-direction, producing displacement into the foreland.

The limb between the two folds is offset by the imbricate thrust with the greatest displacement (B, Figure 3). Units correlate across this poorly exposed fault, but only the hanging wall has cutoffs. Correlating the top of the massive coral-stromatoporoid calcarenites across the fault, the minimum displacement would be 28 m (b' to b", Figure 3). This distance would be the minimum between the hanging wall cutoff (b', Figure 3) and the first possible footwall cutoff (b", Figure 3) The relative age of the thrust to the folds can not be determined, because of the poor exposure and the position in the limb with minimal fold curvature.

The Hopeville anticline (C, Figure 3) has a more complex geometry and style than the syncline, resulting from transection of the anticlinal core by the thrust system. The anticline in the thrust footwall is an upright-subhorizontal, subrounded, tight Class IC-to-IB fold with small disharmonic parasitic folds. However, the anticline in the hanging wall has a box shape with subrounded-to subangular hinges.

The thrust trajectory changes across the anticline. The thrust is horizontal to southeasterly-dipping and bedding-paralled in the northeastern limb (A, Figure 4). Through the core, the thrust changes to horizontal,

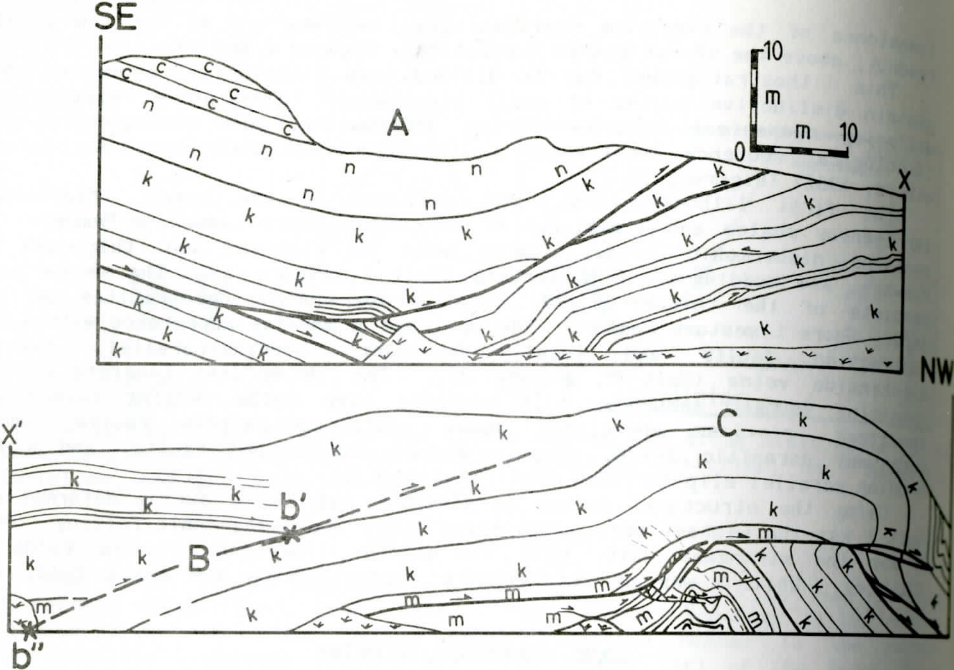

Figure 3. Structural profile of roadcut, using lithostratigraphy. Lithological symbology as in Figure 2. X to X' is contiguous. The thin dashed lines in Figure 3 and 4 are prominent joint surfaces. See text for details.

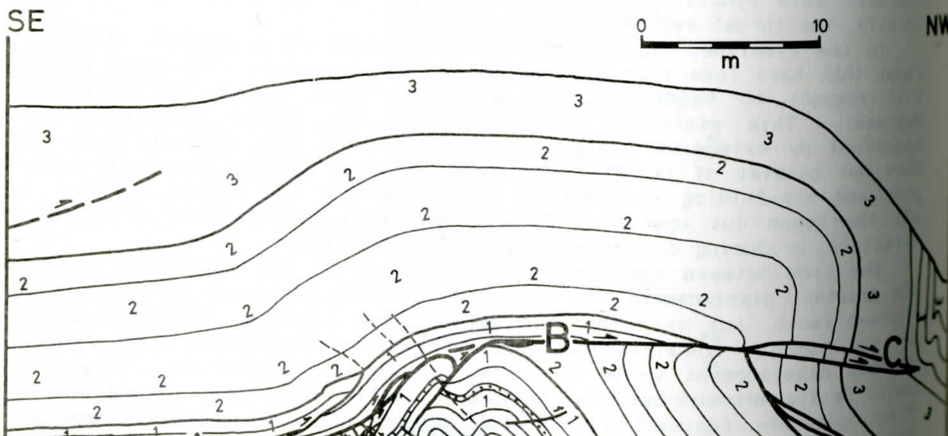

Figure 4. Structural profile of core of Hopeville anticline (position C in Figure 3), using lithostructural units.

and cuts upsection to the northwest (B, Figure 4). The thrust splays in the northwestern limb into two imbricates with less than 1 m total displacement (C, Figure 4) and the main thrust with an additional 12 m displacement. The main fault then splays into two northwesterly dipping segments, which are layer-parallel (D, Figure 4) and layer-oblique, cutting up-section to the northwest (E, Figure 4).

Whether the anticline is older or younger than the thrust system is not immediately obvious. However, this information combined with the relative ages from the syncline would provide the overall deformation sequence for the folds and thrust system.

DEFORMATION SEQUENCE IN ANTICLINE

To determine whether the anticline or the thrust system is older, two palinspastic restorations were performed (Figure 5). In the first restoration, the effects of younger thrusting were removed, and in the second, the effects of younger anticlinal folding were removed. For the restorations, simple rotations and translations were applied that preserved bed length, bed thicknesses, and specified angles. However, the restorations do not constitute balanced sections (Dahlstrom, 1969) because pinpoints for comparing lengths of different beds could not be established.

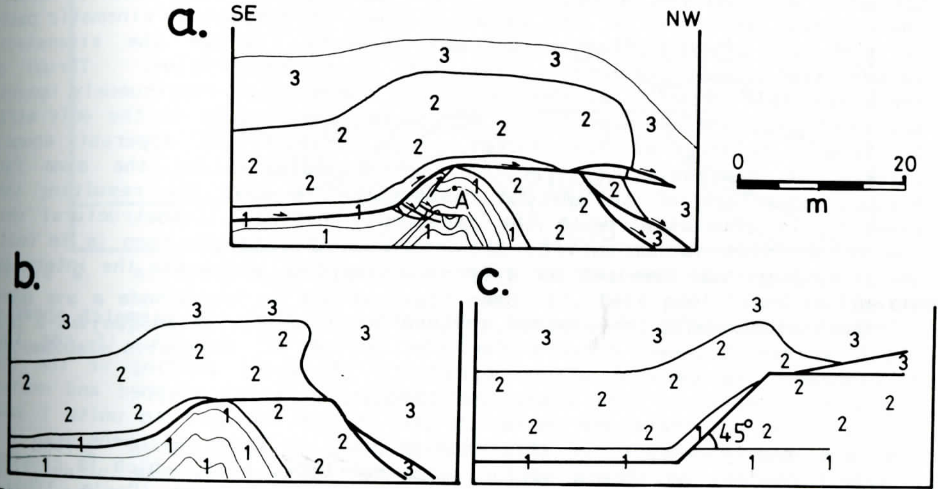

Figure 5. a) Simplified present geometry in Hopeville anticline with lithostructural units. b) Restored geometry if only thrust displacements are removed. c) Restored geometry if only folding is removed.

First Restoration: Anticline Older Than Thrust System

This deformation sequence is preferred by Drabish and Sites (1982), who sketched the northwestern end of the locality (equivalent in position to Figure 4). Their reasons for the sequence choice were 1) subhorizontal trajectory of thrust through the anticline, 2) suggested overturning of the anticline in the hanging wall, and 3) consistency of the southeastern-to-northwestern transport direction with regional transport.

For the restoration in the present study, footwall geometry was preserved, and thrust displacement was removed by matching the footwall and hanging wall cutoffs of the top of lithostructural unit 2 (Figure 5b). This restoration produces several problems. First, to preserve bed lengths and bedding-to-fault angles in the hanging wall, the anticline is a box fold with an unusually bulbous northwestern nose. Second, the fold geometry changes abruptly across the thrust, whereas the two displaced parts of an older fold should form a continuous rematch, if the fault is younger. Third, the footwall and hanging wall cutoffs for the bottom of unit 2 mismatch by about 1 m. Finally, the thrust changes kinematically. In the southeastern limb, reverse-dip slip produces horizontal contraction, but in the northwestern limb, normal-dip slip produces horizontal extension, where the dip direction of the fault reverses.

Second Restoration: Anticline Younger Than Thrust System

For the second restoration, the effects of folding were first removed for the thrust system. Bedding-to-fault angles in the footwall were preserved. Also, beds were rotated about a fold hinge line (A, Figure 5a) to a common level, using a bed in unit 1 as a datum (stippled bed, Figure 4). The resulting thrust geometry is a flat and ramp with a ramp angle that changes from an unusually steep 45 degrees (Boyer and Elliott, 1982) to less than 15 degrees.

Unlike the first restoration, the preservation of bed lengths and bedding-to-fault angles in this restoration produces a simple common geometry (Butler, 1982). Also, in the second restoration, the change in fold style across the thrust can be explained. The change occurs because the anticline deforms a sequence, which had already been locally folded by the rootless anticline of the thrust system. A consistent kinematic pattern is produced by the flat with ramp geometry, unlike the kinematically inconsistent fault geometry after the first restoration. Thrust dip-direction is constantly southeastern; the thrust cuts continuously upsection and up-dip to the northwest; and horizontal contraction is the only strain. Subsequent folding of the thrust would produce the apparent anomalous pattern of horizontal contraction and extension along the same fault. Another advantage of the second restoration is that the resulting thrust geometry is consistent with the previously defined lithostructural units. The thrust flat is in unit 1, the weakest; the steeper ramp is in unit 2, the strongest; and the gentler ramp is in unit 3, which has the intermediate strength.

One problem with the second restoration is that the mismatch apparently still exists between cutoffs for the bottom of unit 2, if the thrust displacement is removed after unfolding. However, folding of the thrust produces a cause for the mismatch. Bedding-parallel stepped and unstepped crystal-fiber slickensides occur in the folded footwall in units 1 and 2, but are mostly absent in the hanging wall. Those slickensides in the footwall display an up-dip motion sense for upper beds, which is consistent for a fold developing with an interlaying-slip component (Davis, 1984). A result of this slip is selective internal rotation of only the footwall beds. The rotation would shorten the footwall ramp with volumetric displacement (Figure 6a), producing the cutoff mismatch. The hanging-wall ramp would be the longer one after folding, which is the case in the anticline.

The localized interlayer-slip would produce several other structures in the footwall (Figure 6b). First, the tighter fold would partially result from filling the volume, vacated by unit 2. The extension faults across the fold core would perform the same function. Second, the bedding in the footwall that is deflected outwards against the fault would result from attempting to preserve the original footwall ramp-length in the stiff unit during rotation. Third, the unusually steep ramp angle would result from the interlayer-slip causing a four to five degree increase in the footwall bedding-to-fault angle (Figure 6a).

Thus, all the problem in the first restoration, where the anticline is the older structure, are solved in the second restoration, where the thrust system is older. Therefore, based on the results of the two restorations, the thrust system is older than the anticline. The geometric complexity in the anticline is the result of folding a thrust ramp with localized interlayer-slip in the footwall. Also, relative ages in the syncline and anticline are the same because the thrust system is consistently older.

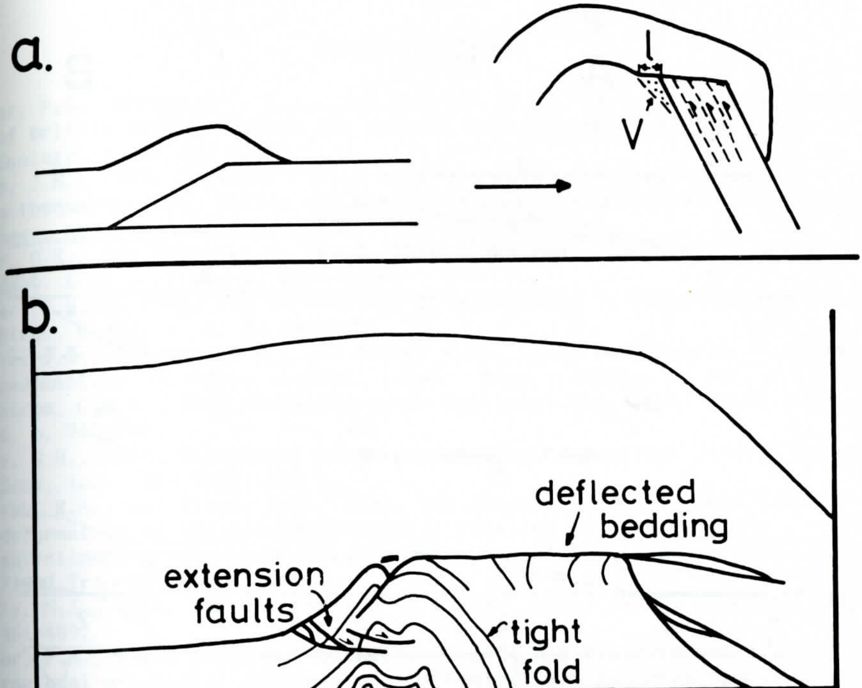

Figure 6. a. Effects of interlayer-slip in footwall during folding of a ramp are a shortening of the footwall ramp (l), here positioned at the base, and a corresponding volume shift (V). b. Structures produced in footwall to accommodate effects.

DISCUSSION

Having established a consistent deformation sequence, a limited palinspatic section may be produced. An artificial pinpoint for the restoration was established in the syncline core by correcting for known thrust displacements. A second pinpoint could not be established, so the section is not completely balanced.

The palinspastic section (Figure 7) is 246 m long parallel to the horizontal, as compared to 187 m for the deformed section. The profile shortening is a minimum of 59 m, or 23.9%. The shortening may be greater because only a minimum displacement may be determined for the major thrust (B, Figure 3). Also, internal distortions by microscopic deformation mechanisms, such as calcite twinning and grain-boundary sliding (but not pressure solution), may have produced up to 10% shortening (Hobbs and others, 1976).

The thrust system produced 16.7% shortening as opposed to only 7.2% for the folds. Thus, the thrust system is not subsidiary to the folds, because 1) the thrusts are older and not coeval, 2) the system is not restricted to a single fold, and 3) the folds shorten less than the thrusts.

Given that the thrust system is older and more important that the parasitic folds, the question still remains as to whether the system is older than Wills Mountain anticline. Two likely possibilities exist. On one hand, the thrust system is older than Wills Mountain anticline and is a part of a larger imbricate thrust system. The larger system would have formed during the emplacement of an older horse of Cambro-Ordovician rocks to the southeast of Wills Mountain anticline. Alternately, the thrust system is coeval with Wills Mountain anticline and is the local precursor

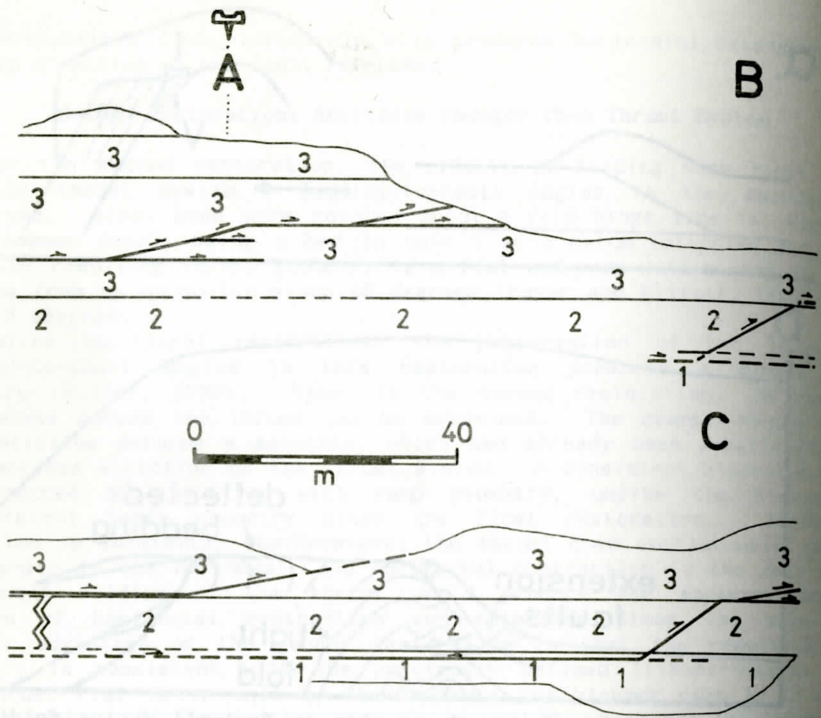

Figure 7. Palinspastic section for Figure 3 with corresponding positions A, B, and C. Artificial pinpoint at syncline core, A.

structure to the folds in the structural terrace. The thrusts would have formed during decoupling of the terrace across the Martinsburg Fortmation.

As the entire extent of the thrust system is unknown, neither possibility may be definitely selected. However, some circumstantial evidence indicates that the first possibility is more likely. First, the thrusts do not cut upsection to the southeast, which would be the uplimb sense for the Wills Mountain anticline (Figure 1). Second, unlike the folds with their upright axial surfaces and symmetric limb-lengths, the thrust system shortens preferentially in only the northwestern direction. Using the fold geometry as an indicator, precursor thrusts in the structural terrace would be expected to have an equal displacement along backthrusts. Third, the thrust system is a continuous network of flats and ramps, rather than isolated contraction faults that would be expected to form as precursors to folds (Berger and others, 1979). Fourth, as Drabish and Sites (1982) noted, the thrust system's northwestern transport direction is consistent with regional transport. Thus, the system may be a small portion of a regional thrust system predating the Wills Mountain anticline.

CONCLUSIONS

1) This mesoscopic thrust system in the Hopeville syncline and anticline is older than the folds, not subsidiary to the folds, and caused greater shortening than the folds.
2) Local complexity in the structural geometry is a result of folding a ramp with differential interlayer-slip in the footwall during the folding.
3) The thrust system probably predates the Wills Mountain anticline and may have formed during the displacement of a thrust horse older than the one beneath Wills Mountain anticline.

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SHOALS JUNCTION AND DUE WEST DOLERITES, SOUTH CAROLINA

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ABSTRACT

The Schoals Junction Dolerite in Abbeville and Greenwood Counties and the Due West Dolerite of Abbeville County, like other western South Carolina dolerites, strike N35°W - N40°W. Field and magnetometer survey data suggest the Shoals Junction Dolerite consists of a swarm of small dikes. Chemical compositions of the dolerites indicate that both are olivine normative but that they belong to separate magmatic lineages. The analyses support division of western South Carolina dolerites into two olivine-normative magma types. The mineralogy of the dikes is consistent with the differences in their chemical composition, with pyroxene compositions being most diagnostic. Pyroxenes in the Shoals Junction Dolerite trend in composition from augite to ferroaugite, whereas in the Due West Dolerite pyroxene compositions trend from augite through subcalcic augite to pigeonite.

INTRODUCTION

Mesozoic dolerites of eastern North America have been the subject of a number of recent geochemical studies directed toward characterization of principal magma types. Weigand and Ragland (1970) and Smith and others (1975) distinguished three main dolerite magma types: olivine-normative, high-TiO_2 quartz-normative, and low-TiO_2 quartz-normative. Weigand and Ragland (1970) also noted an unequal geographic distribution of olivine-normative and quartz-normative magma types. In the Carolinas, for example, olivine-normative dolerites predominate. Thus, all South Carolina dolerite analyses reported by Weigand (1970) are olivine-normative. Later studies (Gottfried and others, 1983; Ragland and Whittington, 1983) indicate the existence of two (at least) independent olivine-normative dolerite magma types, which Ragland and Whittington (1983) have called low-LIL olivine tholeiites (LLO) and high-LIL olivine tholeiites (HLO).

Warner and others (1985) reported chemical data for five dolerite dikes from Pickens and Greenville Counties, western South Carolina. They recognized two independent olivine-normative magma types which roughly correspond to the LLO and HLO groups of Ragland and Whittington (1983). Warner and others (1985) further described some mineralogical criteria which may help to discriminate between the two dolerite magma types. These criteria can be related to differences in the dike compositions.

In this paper we describe the mineralogy and geochemistry of the Shoals Junction and Due West dolerite dikes that were mapped during a study of ground water quality and yield in Greenwood County (Snipes and others, 1984). Our principal goal has been to see how these newly described dikes fit into the classification scheme for western South Carolina dolerites developed by Warner and others (1985).

METHODS OF STUDY

Modes of dolerite samples were determined by point counting; opaque minerals were identified and their abundance estimated using reflected light microscopy on polished 1" diameter microprobe mounts. Mineral compositions were obtained with an automated MAC-400 electron microprobe at NASA, Goddard Space Flight Center. The analyses were made using a beam size of ~ 2 μm at an accelerating potential of 15 kv. The mineral analyses were corrected for differential matrix effects following the procedure of Bence and Albee (1968) and Albee and Ray (1970). Analyses of opaque oxide minerals were recalculated on the basis of stoichiometry to determine Fe_2O_3 and FeO from total Fe. Bulk chemical analyses were made using a Phillips model 1410 manual vacuum X-ray fluorescence spectrometer at City College, City University of New York. Preparation of the samples was according to techniques discussed by Claisse (1956) and Hutchinson (1974). Replicate analyses were made utilizing the dual grinding method (Volborth, 1965).

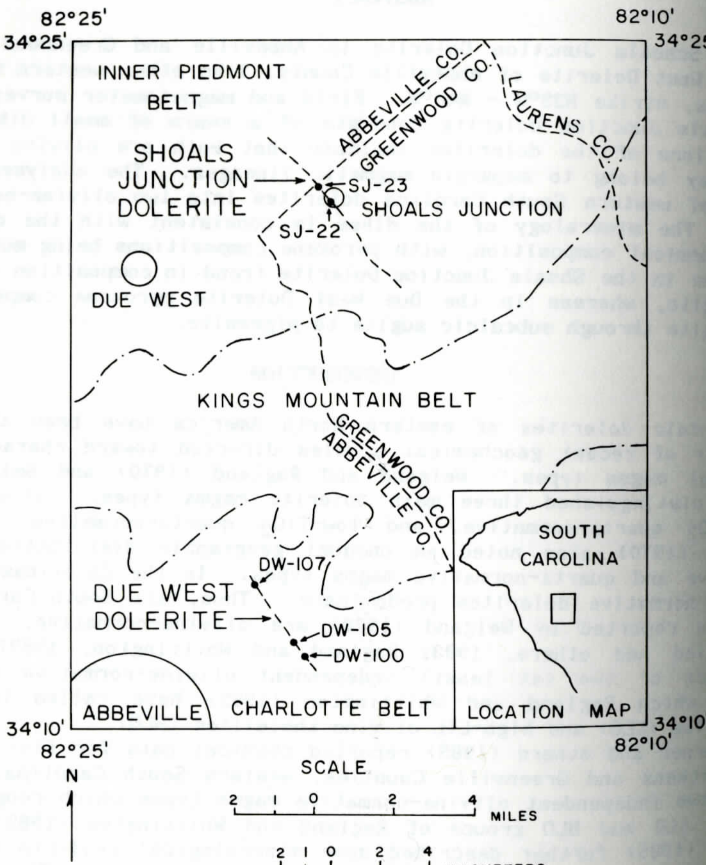

Figure 1. Map of the Due West and Shoals Junction Dolerites. Sample locations are shown by the symbols DW-100, SJ-22, etc. Charlotte, Kings Mountain and Inner Piedmont Belt boundaries are from Overstreet and Bell (1965).

GEOLOGIC SETTING

The locations of the Shoals Junction and Due West Dolerites are shown in Figure 1. The Shoals Junction dike is situated within the Inner Piedmont Belt of Overstreet and Bell (1965) while the Due West intrusive is located within the Charlotte Belt. Elsewhere in South Carolina and other localities in eastern North America, studies by Steele and Ragland (1976), Chalcraft (1976), de Boer and Snider (1979) and Snipes and Furr (1979) suggest that the dolerite dikes were emplaced in steeply dipping extensional fractures during Mesozoic time, perhaps during the Late Triassic or Early Jurassic.

The Due West Dolerite of Abbeville County is situated about 16 km SSE of the town of Due West. Its length is approximately 4.5 km, while its width varies from about 200 m to less than 1 m. The strike is about N35°W, and the dip is essentially vertical. Because of vegetative cover outcrops are few in number and occur in streams or along ridges. Consequently, only sample DW-105 was collected in place, on the margin of a ~ 2 m wide outcrop; samples at localities DW-100 and DW-107 (Figure 1) were collected from the middle of a field of dolerite float.

The length of the Shoals Junction Dolerite is approximately 8.7 km. Its northern end is in Abbeville County, about 2.9 km north of the county line, while it terminates in Greenwood County, about 5.8 km south of the county line. Its strike varies from N32°W to N50°W, but the dominant trend is about N40°W. Its dip is vertical or nearly so. As is the case with the Due West Dolerite, actual outcrops are rare. For the most part the dolerite outcrops are 10 m to less than 1 m in width; however, cobbles and boulders of dolerite float are common and occasionally such areas are more than 250 m wide.

At one locality (lat. 34°22'25" N and long. 82°19'35"W) the Shoals Junction Dolerite consists of three small vertical dikes, each having a width of about 0.3 m (Snipes and others, 1984). The dikes intrude granitic material and are spaced about 14 m apart. Moreover, scattered cobbles and boulders of dolerite float occur to the northeast, perpendicular to the stike of the doleritic rocks, for a distance of about 100 [illegible] The float is found in a light colored, sandy soil which we feel was derived from the weathering of granitic rocks. The occurrence of predominantly light colored soil in association with dolerite float has been observed at several other localities, for example, along a ridge (lat. 34°21'37"N and long. 82°18'37" W) situated 2 km south of the foregoing locality. Elsewhere in the Piedmont, where the rocks are mainly doleritic, the soils are predominantly dark red.

The magnetic profile of the Shoals Junction Dolerite is much more irregular than those of the Due West and Easley Dolerites (Snipes and others, 1984). Moreover, its maximum magnetic intensity, 36×10^3 gammas, is considerably less than those of the Due West and Easley dikes, which are 78×10^3 and 76×10^3 gammas respectively. This magnetometer data together with field observations indicate that the Shoals Junction Dolerite consists of a swarm of small dikes, which were probably emplaced in fractures of either faults or joints. Pebbles and cobbles of quartz mylonite float are commonly associated with the dolerite, but we have not been able to find a well-exposed contact between the dolerite and the mylonite in order to determine the relationship between the two.

DIKE COMPOSITIONS

Major element abundances for the Shoals Junction and Due West Dolerites are reported in Table 1. The two samples of the Shoals Junction Dolerite (see Figure 1) are very similar in composition. Likewise the three samples of the Due West Dolerite show little variation in composition. However, there are some significant compositional differences between the two dolerites. Most notably, K_2O abundance is about an order of magnitude less in the Shoals Junction Dolerite (0.05 - 0.07 wt. %) than in the Due West Dolerite (0.43 - 0.45 wt. %). The abundances of SiO_2, P_2O_5 and TiO_2 are less in the Shoals Junction Dolerite than in the Due West Dolerite, whereas CaO, Al_2O_3 and Na_2O occur in slightly greater amounts in the Shoals Junction Dolerite than in the Due West samples. Fe_2O_3 and MgO concentrations are about the same in the two dikes and, consequently, their mafic indices, 100X $Fe_2O_3/(Fe_2O_3 + MgO)$, are similar.

CIPW norms for the dolerites are also shown in Table 1. Because all Fe was determined as Fe_2O_3, an assumption had to be made about the relative proportions of FeO and Fe_2O_3 in order to compute the normative mineralogy. Based on the median composition of olivine-normative (Quarryville type) diabase reported by Smith and others (1975), we assumed a ferric/ferrous iron ratio of 0.15 and calculated FeO and Fe_2O_3 accordingly (a ratio of about 0.15 seems reasonable from the mineralogy of the dolerites, also). Both dolerites are clearly olivine-normative. However, there are major differences in the normative mafic mineralogy of the two dikes. In the Shoals Junction Dolerite normative olivine and pyroxene are nearly equal, whereas the norms computed for the Due West Dolerite yield 2 - 3 times as much pyroxene as olivine. Further, the amount of normative hypersthene is much less in the Shoals Junction Dolerite.

Table 1. Major oxide composition (wt. %) and CIPW norms of dolerites

	Shoals Junction		Due West		
	SJ-22	SJ-23	DW-100	DW-105	DW-107
SiO_2	46.93	47.28	47.91	48.29	48.65
TiO_2	0.68	0.58	0.76	0.73	0.75
Al_2O_3	15.30	15.54	14.53	14.61	15.00
Fe_2O_3*	13.80	12.61	12.87	12.49	12.32
MgO	9.95	10.48	10.35	9.73	9.78
CaO	11.48	11.65	10.20	10.70	10.26
Na_2O	2.15	2.27	2.11	2.02	2.11
K_2O	0.07	0.05	0.43	0.45	0.43
P_2O_5	0.03	0.02	0.07	0.08	0.07
Total	100.39	100.48	99.23	99.10	99.37
#MI	58	55	55	56	56
		CIPW norms			
Q	---	---	---	---	---
Or	0.4	0.3	2.6	2.7	2.5
Ab	18.2	19.2	17.9	17.1	17.9
An	31.9	32.1	28.9	29.5	30.2
Di	20.3	20.8	17.3	18.8	16.5
Hy	4.2	1.7	12.6	14.6	16.7
Ol	19.8	21.4	14.6	11.1	10.4
Mt	3.0	2.7	2.8	2.7	2.7
Il	1.3	1.1	1.4	1.4	1.4
Ap	0.1	0.1	0.2	0.2	0.2

*All Fe determined as Fe_2O_3
#MI, mafic index = 100X $Fe_2O_3/(Fe_2O_3 + MgO)$

PETROGRAPHY AND MINERAL CHEMISTRY

Modal abundances of the dolerite samples are presented in Table 2. The chief minerals are plagioclase (bytownite - labradorite), pyroxene (principally augite), and olivine. The Shoals Junction Dolerite contains more olivine and plagioclase and considerably less pyroxene than does the Due West Dolerite. The ratio of modal pyroxene/olivine is thus greater in the Due West Dolerite, which is consistent with the normative mineralogy (Table 1). Other constituents present are opaque minerals (mostly titanomagnetite with minor sulfides and, in the Due West samples only, some chromite), various alteration products, including chlorite and iddingsite derived from alteration mostly of olivine, and small amounts of brown, glassy to cryptocrystalline mesostasis. One sample (DW-107) of the Due West Dolerite also contains some biotite.

Textures of the dolerites are illustrated in Figure 2. Thin sections of the Shoals Junction Dolerite exhibit an intergranular texture (Figures 2A, B) with equant to subequant, euhedral to subhedral olivine crystals and

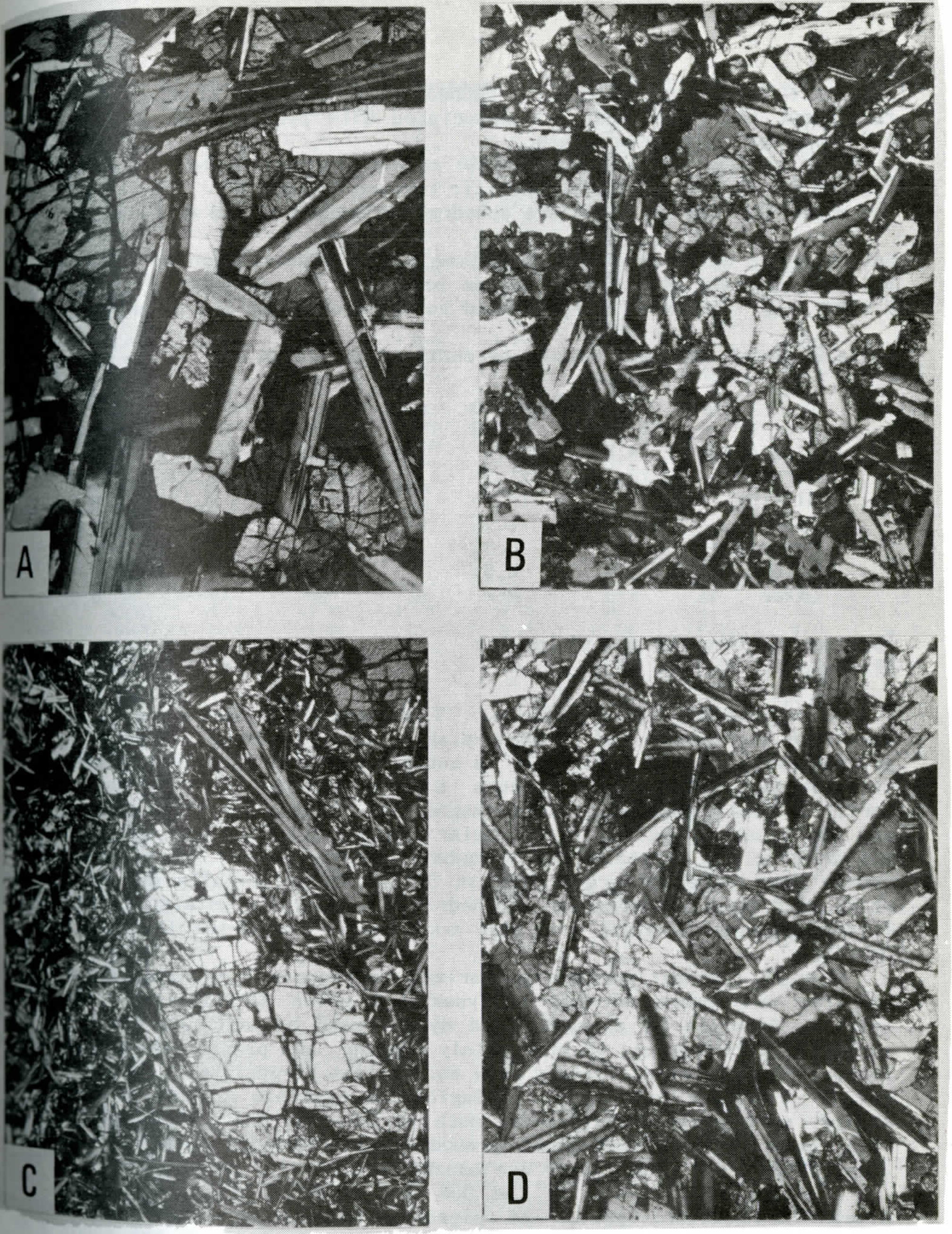

Figure 2. Photomicrographs (transmitted light, X nicols) illustrating dolerite textures. Long dimension equals about 2 mm. A. SJ-22: subsequant olivine and augite crystals occupy interstices between plagioclase laths. B. SJ-23: texture similar to SJ-22 but finer-grained. C. DW-100: olivine microphenocryst surrounded by plagioclase - augite matrix. D. DW-107: subophitic intergrowth of plagioclase with augite.

subequant to irregular, anhedral pyroxene grains surrounded by plagioclase laths. Much of the olivine occurs as 0.5 - 1 mm crystals, but there are also abundant small (0.05 - 0.1 mm across) olivine crystals. The plagioclase laths range in length from less than 0.5 mm to nearly 3 mm. Titanomagnetite occurs as irregular, subhedral or anhedral crystals disseminated throughout the rock; most are associated with pyroxene and plagioclase but a few contact olivine crystals.

Two samples (DW-100 and DW-105) of the Due West Dolerite are very fine-grained and display a microporphyritic texture (Figure 2C). The microphenocrysts are primarily olivine; they are euhedral to subhedral and often have either hollow cores or embayed margins. Some plagioclase is also microphenocrystic, but the majority of the plagioclase occurs in the groundmass as laths 0.2 - 0.3 mm long. Pyroxene is restricted to the groundmass and is present as

Table 2. Modal analyses (volume percent) of dolerites

	Shoals Junction		Due West		
	SJ-22	SJ-23	DW-100	DW-105	DW-107
Plagioclase	54.6	54.4	46.2	46.2	49.3
Pyroxene	22.1	21.9	38.1	38.5	33.8
Olivine	19.2	17.3	7.1	7.1	8.5
Opaques	1.5	1.7	1.9	2.3	2.0
Mesostasis	0.7	0.3	1.3	1.1	0.6
Alteration*	1.8	4.5	5.5	4.9	4.9
Other	---	---	---	---	0.9#

*Dominantly chlorite and iddingsite
#Principally biotite

< 0.2 mm subequant intergranules. Titanomagnetite occurs as tiny crystals scattered throughout the pyroxene-plagioclase groundmass. Chromite crystals are generally included within or in contact with the olivine microphenocrysts; an occasional chromite grain occurs in the groundmass. The latter are usually partly or wholly mantled by titanomagnetite. Sample DW-107 is coarser-grained and has a subophitic to intergranular texture (Figure 2D). Although olivine and plagioclase crystals are somewhat larger than in the microporphyritic rocks, the increased grain size in DW-107 is principally due to pyroxene, which is present as irregular, subhedral or anhedral crystals as large as 1 mm. Titanomagnetite crystals are also much larger in DW-107 than in DW-100 or DW-105.

Mineral compositions are summarized in Figures 3 - 5 and representative electron microprobe mineral analyses are given in Table 3. The Shoals Junction Dolerite is characterized by pyroxene compositions that trend from augite to ferroaugite, olivine mainly Fo_{80-85}, and plagioclase mostly An_{70-80} (Figure 3). Olivine crystals show only slight Fe-enrichment (generally a few mol %) from core to margins. Plagioclase crystals display more pronounced zoning, with crystal margins as much as 30 - 40 mol % more sodic than the lath cores. A number of small groundmass crystals more sodic than An_{50} also occur (Figure 3). The latter are usually situated in mesostasis-rich areas.

Pyroxene compositions in the Due West Dolerite trend from augite toward pigeonite; however, grains of pigeonite composition were found only in DW-107 (Figure 4). Pyroxene compositions in this sample are about 7 or 8 mol % more Mg-rich than are pyroxenes in the two finer-grained, microporphyritic samples. Olivine compositions are Fo_{81-86} in the microporphyritic samples but are for the most part distinctly more Fe-rich (Fo_{57-68}) in DW-107 (Figure 4). There are a few Mg-rich olivines (about Fo_{85}) in the latter sample, however. In both microporphyritic and subophitic samples the olivine grains show minor compositional zonation, with crystal margins typically a few mol % richer in fayalite component than grain cores. Unlike the mafic minerals, plagioclase compositions in all samples of the Due West Dolerite are similar; most laths are An_{66-73} but there are some crystals of more sodic composition (Figure 4). As in the Shoals Junction Dolerite, the more sodic plagioclases are small laths in mesostasis-rich areas. Many plagio-

Table 3. Electron microprobe analyses (wt. %) of minerals in Shoals Junction and Due West Dolerites

	1	2	3	4	5	6	7	8	9
SiO_2	49.04	60.06	50.14	54.49	50.90	39.71	36.64	39.61	35.61
Al_2O_3	32.68	24.59	32.13	28.92	31.82	na	na	na	na
FeO	na	na	na	na	na	15.80	28.72	14.85	35.74
MgO	na	na	na	na	na	43.51	33.02	45.21	28.35
CaO	15.01	6.45	14.46	11.50	13.82	0.38	0.58	0.39	0.44
Na_2O	2.73	7.40	2.94	4.52	3.28	na	na	na	na
K_2O	0.03	0.35	0.09	0.27	0.16	na	na	na	na
Total	99.48	98.86	99.76	99.70	99.98	99.40	98.96	100.06	100.14
Number of Oxygens =	32	32	32	32	32	4	4	4	4
Si	8.985	10.796	9.142	9.848	9.246	1.005	0.996	0.993	0.990
Al	7.055	5.206	6.900	6.157	6.810	--	--	--	--
Fe	--	--	--	--	--	0.334	0.652	0.311	0.831
Mg	--	--	--	--	--	1.642	1.336	1.690	1.175
Ca	2.946	1.239	2.819	2.222	2.686	0.009	0.017	0.010	0.012
Na	0.968	2.579	1.038	1.580	1.151	--	--	--	--
K	0.005	0.079	0.021	0.058	0.037	--	--	--	--
Sum	19.959	19.899	19.919	19.865	19.929	2.991	3.000	3.004	3.007
An	75.2	31.8	72.7	57.6	69.3				
Ab	24.7	66.2	26.8	40.9	29.7				
Or	0.1	2.0	0.5	1.5	1.0				
Fo						83.1	67.2	84.5	58.6
Fa						16.9	32.8	15.5	41.4

	10	11	12	13	14	15	16	17	18
SiO_2	50.52	49.75	49.70	51.29	52.82	na	na	na	na
TiO_2	0.48	0.46	1.05	0.71	0.27	16.19	17.33	0.31	0.56
Al_2O_3	3.18	1.12	5.27	2.41	2.00	3.03	1.71	37.59	32.17
Cr_2O_3	0.20	na	0.17	na	na	0.02	0.20	25.59	24.21
*Fe_2O_3	na	na	na	na	na	34.20	33.22	6.35	10.85
*FeO	11.91	23.75	10.81	16.47	21.00	45.44	45.42	14.17	21.46
MnO	0.25	na	0.21	na	na	0.92	1.17	0.23	0.29
MgO	14.46	9.46	13.31	16.87	21.29	0.15	0.46	15.48	10.25
CaO	18.84	15.67	19.60	12.74	4.74	na	na	na	na
Na_2O	0.28	0.17	0.29	0.25	0.13	na	na	na	na
Total	100.11	100.38	100.41	100.75	102.23	99.95	99.51	99.72	99.79
Number of Oxygens =	6	6	6	6	6	4	4	4	4
Si	1.896	1.948	1.853	1.915	1.932	--	--	--	--
Ti	0.013	0.012	0.029	0.020	0.006	0.454	0.490	0.007	0.013
Al	0.140	0.051	0.230	0.105	0.085	0.133	0.076	1.270	1.148
Cr	0.005	--	0.004	--	--	0.001	0.006	0.580	0.579
Fe^{3+}	---	--	---	--	--	0.959	0.939	0.137	0.247
Fe^{2+}	0.373	0.777	0.337	0.514	0.642	1.416	1.427	0.340	0.543
Mn	0.007	--	0.005	--	--	0.029	0.037	0.006	0.007
Mg	0.808	0.552	0.739	0.938	1.160	0.008	0.026	0.661	0.462
Ca	0.756	0.657	0.782	0.509	0.186	--	--	--	--
Na	0.020	0.012	0.021	0.017	0.008	--	--	--	--
Sum	4.019	4.010	4.001	4.017	4.019	3.000	3.001	3.001	2.999
Wo	39.0	33.1	42.1	26.0	9.4				
En	41.7	27.8	39.8	47.8	58.4				
Fs	19.3	39.1	18.1	26.4	32.3				

na = not analyzed
*Fe_2O_3 and FeO in spinel group minerals calculated on the basis of stoichiometry
Analyses: 1 - plagioclase, SJ-23; 2 - sodic plagioclase, SJ-23; 3, 4 - core and rim of zoned plagioclase, DW-105; 5 - plagioclase, DW-107; 6,7 - core and rim of zoned olivine, SJ-22; 8 - olivine, DW-105; 9 - olivine, DW-107; 10 - augite, SJ-22; 11 - ferroaugite, SJ-23; 12 - augite, DW-105; 13 - subcalcic augite, DW-107; 14 - pigeonite, DW-107; 15 - titanomagnetite, SJ-23; 16 - titanomagnetite, DW-107; 17 - chromite, DW-105; 18 - chromite, DW-107

clase laths are normally zoned with margins as much as 15 - 25 mol % more sodic than the central portions of the laths. The compositions of chromite crystals in the Due West Dolerite are indicated in Figure 5. Chromites in the microporphyritic-textured samples are more Mg-rich than those in DW-107. The ratio Cr/Cr + Al is close to 0.3, and there is a distinct trend whereby Cr/Cr + Al increases slightly with decreasing Mg/Mg + Mn + Fe^{2+}.

DISCUSSION

Warner and others (1985) found that Mesozoic dolerite dikes of western South Carolina can be assigned to two different olivine-normative magma types. These are hereafter referred to as the Fews Chapel and Cleveland magma types based on informal dike names proposed by the above-mentioned authors. The Fews Chapel magma type is characterized by higher Al_2O_3, Fe_2O_3 and Na_2O and lower SiO_2, MgO and K_2O compared to the Cleveland magma type. Based on the dike compositions reported in Table 1, we classify the Shoals Junction Dolerite with the Fews Chapel magma type and the Due West Dolerite with the Cleveland group. The Shoals Junction Dolerite is virtually iden-

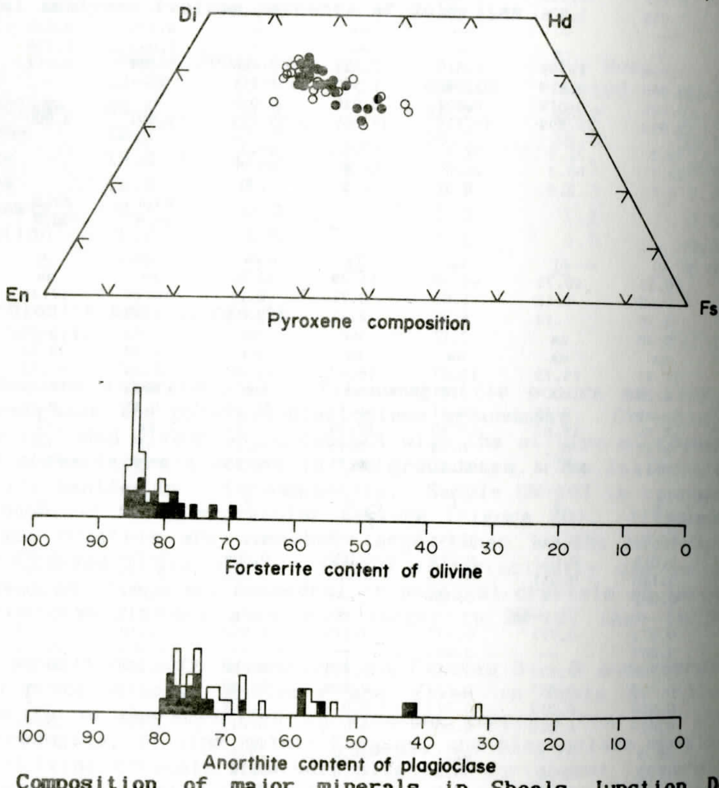

Figure 3. Composition of major minerals in Shoals Junction Dolerite. Pyroxene compositions plotted on pyroxene quadrilateral. Olivine and plagioclase compositions summarized in histograms. Symbols: filled, SJ-22; open, SJ-23.

tical in major oxide composition with the Fews Chapel Dolerite, except for K_2O, which is lower (0.05 - 0.07 wt. % in Shoals Junction vs. 0.22 - 0.27 wt. % in Fews Chapel). The Due West Dolerite contains higher Fe_2O_3 and lower MgO but otherwise is compositionally similar to analyses of the Cleveland magma type. Because of the higher Fe_2O_3 coupled with lower MgO, the Due West samples have higher mafic indices than samples of the Cleveland-type dolerites (55 - 56 vs 47 - 49, respectively). This implies that the Due West dike may represent Cleveland-type magma that had undergone some differentiation prior to or during emplacement.

Mineral modal abundances are also consistent with assignment of the Shoals Junction Dolerite to the Fews Chapel magma type and the Due West Dolerite to the Cleveland magma type. For example, Warner and others (1985) found that samples of Fews Chapel-type dolerite contain about 10 volume %

more modal plagioclase and have a lower modal pyroxene/olivine ratio than do samples of Cleveland-type dolerite. Data from Table 2 indicate a similar relationship holds for the Shoals Junction and Due West Dolerite samples. Another significant modal parameter noted by Warner and others (1985) is chromite abundance: the Cleveland-type dolerite samples contain 0.2 - 0.3 volume %, whereas chromite is only a trace constituent in Fews Chapel-type dolerites. Samples of the Due West Dolerite contain 0.1 - 0.2 % modal chromite, but no chromite was detected at all in samples of the Shoals Junction Dolerite.

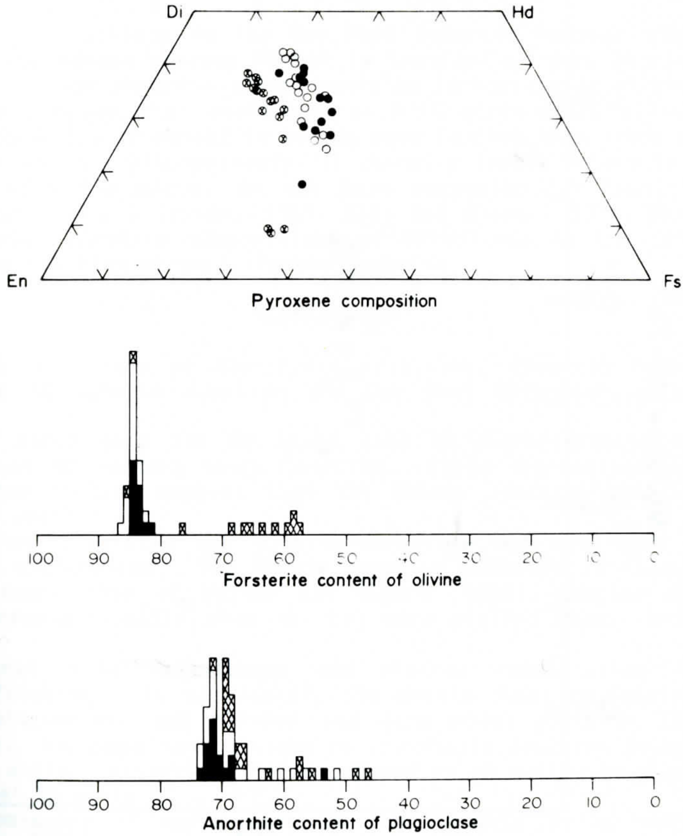

Figure 4. Composition of major minerals in Due West Dolerite. Symbols: filled, DW-100; open, DW-105; cross-hatched, DW-107.

Pyroxene compositions in the Shoals Junction and Due West Dolerites follow markedly different trends. In the Shoals Junction Dolerite they trend from augite to ferroaugite (Figure 3), whereas pyroxene compositions in the Due West dike plot more or less along a trend extending from augite through subcalcic augite to pigeonite (Figure 4). The augite → ferroaugite trend in the Shoals Junction Dolerite is very similar to the trend observed in the Fews Chapel Dolerite, while the augite → pigeonite trend for the Due West Dolerite closely parallels the pyroxene trend displayed by Cleveland-type dolerites (Warner and others, 1985). We note that the lack of Ca-poor pyroxene in the Shoals Junction Dolerite samples correlates with the CIPW norms, which show much lower normative hypersthene in these rocks than in samples of the Due West Dolerite (Table 1). Lower hypersthene in the norms reflects lower SiO_2, which causes more Mg and Fe to be assigned to olivine and less to hypersthene in the calculation procedure. The Shoals Junction

Dolerite contains less SiO_2 than the Due West Dolerite (Table 1), and, consequently, as the Shoals Junction Dolerite magma cooled more olivine crystallized from it, depleting the melt in hypersthene component, than in the corresponding case of the Due West Dolerite magma. Hence, the greater modal olivine abundance and the augite → ferroaugite pyroxene trend of the Shoals Junction Dolerite as opposed to the augite → pigeonite trend of the Due West Dolerite are primarily results of the differences in SiO_2 concentrations in the two magmas.

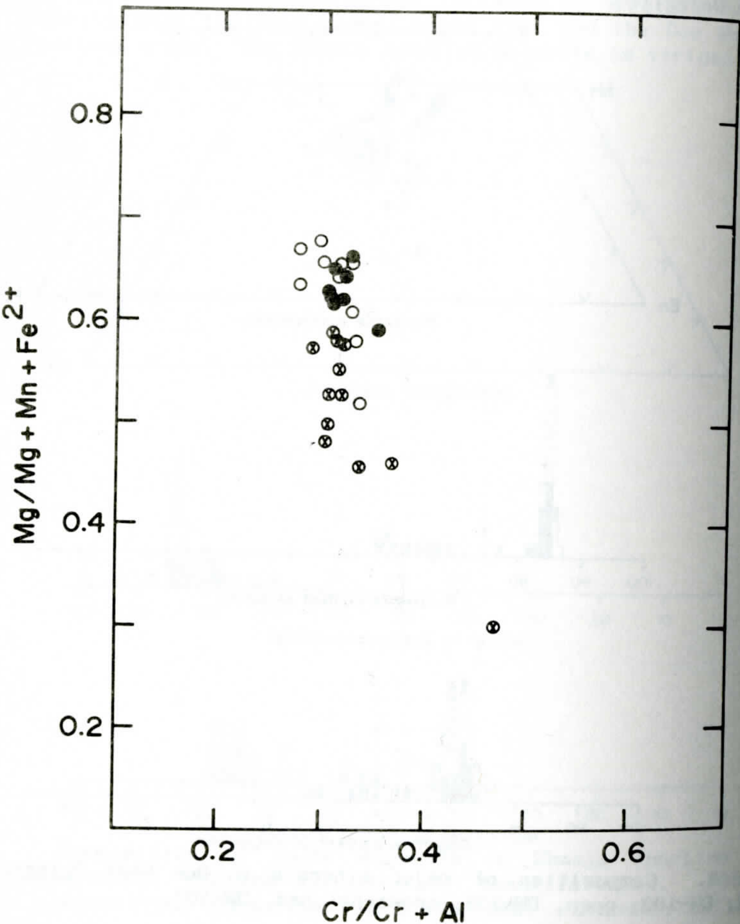

Figure 5. Plot of Mg/Mg + Mn + Fe^{2+} versus Cr/Cr + Al showing chromite compositions in Due West Dolerite. Symbols: same as in Figure 4.

Besides pyroxene, Warner and others (1985) found chromite to be a useful discriminant between the Fews Chapel and Cleveland magma types. Chromite compositions in the Due West Dolerite (Figure 5) overlap those of chromites from samples of Cleveland-type dolerites, but have Cr/Cr + Al ratios higher than is characteristic of Fews Chapel-type dolerites. Thus chromite and pyroxene mineral compositions both are consistent with grouping the Shoals Junction dike with the Fews Chapel magma type and the Due West Dolerite with the Cleveland.

One additional relationship of interest pertains to the differences in mafic mineral compositions (Figure 4) observed among samples of the Due West Dolerite. The coarser-grained, subophitic-textured sample, DW-107 (Figure 2D), is characterized by olivines more Fe-rich and pyroxenes more Mg-rich than occur in the finer-grained, microporphyritic-textured samples, DW-100 and DW-105 (Figure 2C). Presumably DW-100, like DW-105, represents material from near the chilled margin of the Due West Dolerite, while DW-107 is inferred from its texture to have come from the interior of the dike, where cooling would have proceeded at a slower rate. As a consequence of slower cooling, one would expect that olivine reaction with the magma was more extensive in DW-107, causing olivine grains to become more Fe-rich than in the more rapidly cooled microporphyritic samples. More extensive olivine reaction in DW-107 would have the simultaneous effect of locally increasing Mg concentration in the melt, so that pyroxene crystallizing from this melt would be more Mg-rich than in the more rapidly chilled DW-100 and DW-105 samples.

Chromite compositions in the Due West Dolerite samples also exhibit a cooling rate dependence whereby DW-107 is typified by more Fe-rich chromites (Figure 5). Because chromite often occurs as inclusions in olivine, it seems reasonable to suppose that equilibration with surrounding olivine crystals would cause chromites in DW-107 to become more Fe-rich than their counterparts in DW-100 and DW-105. Alternatively, if chromite itself enters into a reaction relationship with the magma, as has been suggested for basaltic rocks by numerous authors (e.g., Irvine, 1967; Hill and Roeder, 1974; Ridley, 1977), the more Fe-rich chromite compositions in DW-107 may be the consequence of more extensive reaction of early-formed chromite.

CONCLUSIONS

Based on our study of the field relations, chemical composition and mineralogy of the Shoals Junction and Due West Dolerites we conclude the following:

1) Both dikes have the NW trend that is characteristic of dolerites in the Piedmont of western South Carolina. Field observations supplemented by magnetometer studies suggest that the Shoals Junction Dolerite consists of a swarm of small dikes.

2) The two dolerites are olivine-normative but have some significant compositional differences. The Shoals Junction Dolerite is classed with the Fews Chapel magma type of Warner and others (1985), whereas the Due West Dolerite is compositionally akin to, but more evolved than, their Cleveland magma type.

3) Mineral modal abundances and mineral compositions support the above classification. In particular, the Shoals Junction Dolerite contains more modal plagioclase and olivine and less modal pyroxene than the Due West Dolerite. Pyroxene compositions in the Shoals Junction Dolerite exhibit a trend from augite to ferroaugite as opposed to an augite to pigeonite trend in the Due West Dolerite.

4) Compositions of mafic minerals and chromite in various samples of the Due West Dolerite appear to be cooling-rate dependent, such that coarser-grained sample DW-107 contains more Fe-rich olivine and chromite and more Mg-rich pyroxene than are found in finer-grained microporphyritic samples.

ACKNOWLEDGMENTS

The authors thank P. J. Wasilewski for making available the automated MAC-400 electron microprobe at NASA's Goddard Space Flight Center for obtaining mineral analyses of the dolerites. The authors also acknowledge S. B. Heaton and D. G. Ries, who assisted with the drafting, and B. Peters, who typed the manuscript. The manuscript has benefitted from critical comments supplied by A. E. Nelson. Financial support for this study was provided by the South Carolina Water Resources Research Institute, Clemson University, and by the U. S. Geological Survey.

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LITHOSTRATIGRAPHY AND LITHOGENESIS OF CONEMAUGH (CARBONIFEROUS) DEPOSITIONAL SYSTEMS NEAR HUNTINGTON, WEST VIRGINIA

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ABSTRACT

The Appalachian Coal Measures have received intensive study, much of it modern. Lacking significant coals or other economially valuable deposits, the Conemaugh Group has not received its proportionate share of this study, and man-made outcrops are generally poor and scattered. Highway construction along both sides of the Ohio and Big Sandy Rivers near Huntington, West Virginia, provides an unparalleled opportunity to study Conemaugh Group rocks. Rocks of the upper Allegheny Group, and the Glenshaw and Casselman Formations of the Conemaugh Group, excellently exposed near Huntington, demonstrate an upward succession from upper delta plain to fluvial environments.

This succession is compatible with previous models (Ferm and others, 1971) in which the Appalachian Carboniferous is interpreted to include a series of clastic wedges prograding north and northwestward with sediments shed by mountains produced by increasing Alleghanian tectonism.

As progradation progressed, the upper delta plain environments represented in the older Carboniferous rocks to the west (Ferm and others, 1976) were replaced in the Conemaugh part of the section by rocks deposited on a lower alluvial plain. These Conemaugh rocks include red claystones, siltstones, and sandstones representing the dominant environments in a fluvial setting: vadose flood plain, natural levee, and channel deposits. Minor lacustrine and paludal deposits can also be recognized.

Two marine invasions (Brush Creek and Cambridge, both Missourian in age) occurred during the deltaic or very early alluvial history of the region, and the last significant marine invasion of eastern North America during the Paleozoic, the Ames, occurred during the Virgilian after fluvial conditions were well established in the area.

INTRODUCTION, GEOLOGIC SETTING, AND PURPOSE

The area of study is located in the tri-state region of Ohio, Kentucky, and West Virginia (Figure 1) and is adjacent to the southwestern termination of the axis of a gently-dipping, doubly-plunging syncline, the Pittsburgh-Parkersburg-Huntington Syncline. Dips in southern Ohio average approximately 5.7 m/km (30 ft/mi) to the southeast toward the syncline axis. Dips along the southeast flank near Huntington locally range up to 7^{o} or 123 m/km (650 ft/mi). The four major lithostratigraphic units, herein considered groups, recognized in the Upper Carboniferous of Ohio (ascending: Pottsville, Allegheny, Conemaugh, and Monongahela) have an aggregate thickness of approximately 1120 ft (340 m) according to Stout (1944). Rocks of the upper Allegheny and Conemaugh Groups encountered in the study area have an aggregate thickness of about 550 ft (168 m) and range in age from late Desmoinesian to at least late Virgilian (Merrill, *in* Lane and others, 1971; 1974).

Parts of this area have been studied previously in some detail but most work has been confined to a single state, for example Condit (1912) in Ohio; Krebs and Teets (1913) and Cross and others, (1956), in West Virginia; Crandall (1877), Dobrovolny and others, (1963), Spencer (1964), Sharps (1967), and Walter (1979) in Kentucky. This study was begun as a mapping project at Louisiana State University in 1965 (Merrill, 1973). It was initiated while I was a graduate student under the direction of J. C. Ferm and additional localities were added over the next decade and more (Merrill,

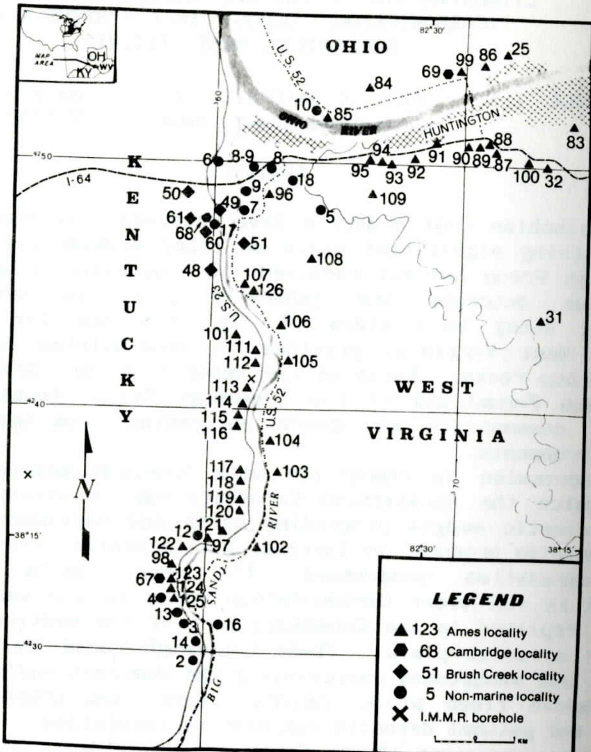


Figure 1. Locality map for measured sections in the Huntington area. United States Geological Survey 7-1/2' quadrangles used as a base for this map, clockwise from the southwest are: Prichard, Burnaugh, Catlettsburg, Huntington, and Lavalette.

1981). These additions coincided with extensive highway construction that made impressive roadcuts that in many cases were coincident with the regional structure, permitting study of persistence and interfingering of individual beds on a scale that has not been possible previously.

The major purpose of this study is to determine the basic stratigraphy of Conemaugh and related rocks in the Huntington area, deduce their general depositional environments, and to relate them to the lithogenetic evolution of the Appalachian Carboniferous.

LOCALITIES AND TECHNIQUES

Except for two boreholes, all localities are surface exposures. With few exceptions these are roadcuts along the principal highways: U.S. 23 in Kewntucky, U.S. 52 in West Virginia and Ohio, and Interstate 64 in West Virginia and Kentucky (Figure 1). Of a total of 68 sections measured in the area, 46 contain the Ames Member, 5 the Brush Creek Member, two contain both the Ames and Cambridge Members, one contains the Brush Creek and Cambridge Members, and 14 contain no marine beds. The locations of these measured sections are shown in Figure 1, their relationship to named units is shown in Figure 2, their Correlations in Figs. 3-5 and their precise locations are given in Table 1.

Sections were measured by hand level and rule and supplemented by field sketches of lateral changes within exposures. Additional field data included photo-mosaics of all important roadcuts along Interstate 64 and U.S. 23. These photo-mosaics not only permitted correlation and

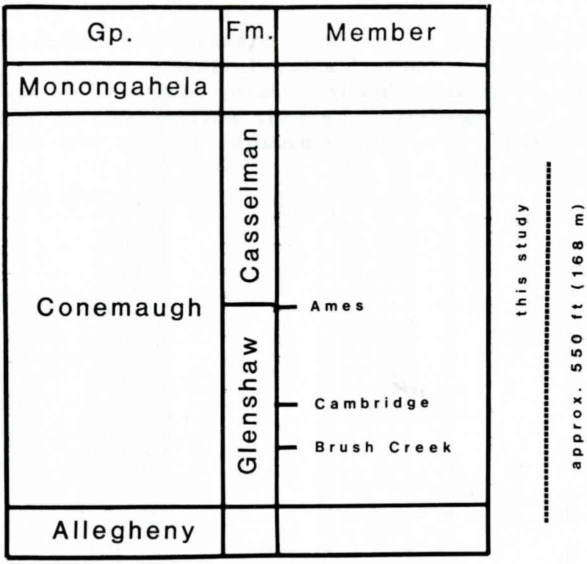


Figure 2. Generalized lithostratigraphic classification and interval studied showing members containing marine fossils. The Missourian-Virgilian chronostratigraphic boundary falls somewhere within the non-marine interval separating the Cambridge and Ames Members.

identification of rock units, but commonly were used to determine the kinds of lateral change, whether interfingering or intergrading, or channeling.

All measured sections were recorded on standard forms provided by the Ohio Division of Geological Survey (address: Fountain Square, Columbus, Ohio 43224) and are on open file there under numbers 10341, 15059, 15060, and 16625 through 16691 (Table 1). Thicknesses and lithologies for each measured section were encoded and these data were used for machine plotting on two different vertical scales in the manner described by Melton and others (1979) and Ferm and others (1979a).

The primary location system employed in this study is the Universal Transverse Mercator (UTM) Grid supplemented with Geographic Coordinates (Table 1). Either system should have precision limits in the 10-20 m (33-66 ft) range. Horizontal distances will be rendered in metres and kilometres with feet and miles parenthetically. Vertical distances were taken and will be reported in Imperial (English) units with SI (metric) units parenthetically.

STRATIGRAPHY AND PETROGRAPHY

Rocks included in this study belong to the uppermost part of the Allegheny Group in West Virginia and the equivalent portion of the Breathitt Formation in Kentucky, and to the lower two-thirds to three-fourths of the Conemaugh Group (same proper name in all three states). This portion of the Conemaugh would embrace all the Glenshaw Formation and the lower part of the Casselman Formation as used in Pennsylvania (Donahue and Rollins, 1974). The base of the Conemaugh in the Huntington region has been drawn on the top of a fairly thick coal bed that corresponds in position and concept with the Upper Freeport in Pennsylvania.

There are three volumetrically dominate rock types within the section studied; all are terrigenous in origin and were deposited under non-marine conditions. In order of abundance they are silt-shales, sandstones (some rudaceous), and "red bed" claystones. The lower part of the section is dominated by sandstones with interbedded silt-shales and thin coals. Higher

Table 1. Locality register. States are abbreviated per postal abbreviations. Other abbreviations, in order of appearance, counties; Lawrence, Wayne, Cabell, Boyd; townships/districts; Union, Guyandotte, Fayette, Westmoreland, Ceredo, Butler; quandrangles; Huntington, Lavalette, Catlettsburg, Prichard, Burnaugh; roads; old U.S. 52, U.S. 52, Interstate 64, old U.S. 23, U.S. 23.

AMES LOCALITIES

loc	GSU#	ST	CO	TP/DIS	QU	RD	UTM coord(17SLN)	N latitude°'"	W longitude°'"
25	15059	OH	LA	UN	HU	X52	71835415	38 25 43	82 28 06
31	16625	WV	WA	UN	LA	-	73054355	38 19 59	82 27 11
32	16626	WV	CA	GU	HU	-	73335038	38 23 41	82 27 09
83	16627	WV	CA	GU	HU	-	74455175	38 24 26	82 26 17
84	10341	OH	LA	FA	CA	-	66105288	38 24 58	82 32 02
85	15060	OH	LA	FA	CA	52	64325211	38 24 33	82 33 15
86	16628	OH	LA	UN	HU	X52	70925390	38 25 33	82 28 44
87	16629	WV	CA	GU	HU	I64	71295068	38 23 46	82 28 26
88	16630	WV	CA	GU	HU	I64	71055100	38 24 00	82 28 42
89	16631	WV	CA	GU	HU	I64	70685117	38 24 05	82 28 52
90	16632	WV	CA	GU	HU	I64	70295105	38 24 02	82 29 09
91	16633	WV	WA	WE	CA	I64	68755078	38 23 52	82 30 10
92	16634	WV	WA	WE	CA	I64	67935060	38 23 45	82 30 44
93	16635	WV	WA	WE	CA	I64	67145060	38 23 41	82 31 18
94	16636	WV	WA	WE	CA	I64	66725047	38 23 40	82 31 35
95	16637	WV	WA	WE	CA	I64	66305048	38 23 39	82 31 51
96	16638	WV	WA	CE	CA	52	61894847	38 22 33	82 34 42
97	16639	KY	BO	-	PR	X23	59703435	38 14 54	82 36 34
98	16640	KY	LA	-	PR	23	58813334	38 14 20	82 36 47
99	16642	OH	LA	UN	HU	52	69535370	38 25 36	82 29 38
100	16643	WV	CA	GU	HU	164	72825018	38 23 39	82 27 22
101	16644	KY	BO	-	BU	-	61044283	38 19 30	82 35 22
102	16645	WV	WA	BU	PR	52	61633406	38 14 46	82 34 51
103	16646	WV	WA	CE	BU	52	62503705	38 16 24	82 34 18
104	16647	WV	WA	CE	BU	52	62133806	38 17 14	82 34 34
105	16648	WV	WA	CE	BU	52	62784168	38 18 52	82 34 10
106	16649	WV	WA	CE	BU	52	62504313	38 19 41	82 34 22
107	16650	WV	WA	CE	BU	52	61184471	38 20 31	82 35 18
108	16651	WV	WA	CE	BU	-	63814615	38 21 18	82 33 31
109	16652	WV	WA	WE	CA	-	66284860	38 22 39	82 31 51
111	16653	KY	BO	-	BU	23	62044235	38 19 15	82 34 40
112	16654	KY	BO	-	BU	23	62034179	38 18 57	82 34 40
113	16655	KY	BO	-	BU	23	61674044	38 18 12	82 34 55
114	16656	KY	BO	-	BU	23	61433968	38 17 52	82 35 03
115	16657	KY	BO	-	BU	23	61403931	38 17 41	82 35 05
116	16658	KY	BO	-	BU	23	61393928	38 17 35	82 35 05
117	16659	KY	BO	-	BU	23	61513736	38 16 33	82 34 59
118	16660	KY	BO	-	BU	23	61553673	38 16 12	82 34 57
119	16661	KY	BO	-	BU	23	61463620	38 15 54	82 35 00
120	16662	KY	BO	-	BU	23	61343563	38 15 36	82 35 09
121	16663	KY	BO	-	BU	23	60763477	38 15 06	82 35 28
122	16664	KY	LA	-	PR	23	59193420	38 14 48	82 36 33
123	16665	KY	LA	-	PR	23	58633280	38 14 03	82 36 55
124	16666	KY	LA	-	PR	23	58693236	38 14 49	82 36 53
125	16667	KY	LA	-	PR	23	58803220	38 13 44	82 36 47
126	16668	WV	WA	CE	BU	52	61274467	38 20 29	82 35 15

CAMBRIDGE LOCALITIES

67	same locality as Ames 123
68	same locality as Brush Creek 61
69	same locality as Ames 99

BRUSH CREEK LOCALITIES

48	16669	KY	BO	-	BU	23	60134587	38 21 07	82 36 00
49	16670	KY	BO	-	CA	-	59924798	38 22 17	82 36 16
50	16671	KY	BO	-	CA	164	59434884	38 22 43	82 36 34
51	16672	WV	WA	CE	BU	52	60894665	38 21 33	82 35 31
60	16673	KY	BO	-	BU	-	60024712	38 21 48	82 36 08
61	16674	KY	BO	-	BU	-	59874763	38 22 05	82 36 14

NON-MARINE LOCALITIES

1	16675	KY	LA	-	PR	23	58272623	38 10 29	82 37 05
2	16676	KY	LA	-	PR	23	59472936	38 12 04	82 36 18
3	16678	KY	LA	-	PR	23	59193138	38 13 17	82 36 29
4	16679	see Ames 125							
5	16680	WV	WA	CE	CA-BU	-	64194802	38 22 20	82 33 17
6	16681	KY	BO	-	CA	23	59845014	38 23 24	82 36 19
7	16682	WV	WA	CE	CA	-	61054808	38 22 20	82 35 25
8-9	16683	WV	WA	CE	CA	I64	61384946	38 23 05	82 35 16
10	16684	OH	LA	FA	CA	52	64095228	38 24 39	82 33 25
12	16685	KY	LA	-	PR	23	59673455	38 14 59	82 36 13
13	16686	KY	LA	-	PR	23	58913163	38 13 26	82 36 43
14	16888	KY	LA	-	PR	23	59573054	38 12 45	82 36 13
16	16689	WV	WA	BU	PR	52	60333099	38 13 05	82 35 44
17	16690	KY	BO	-	BU	-	59914741	38 21 58	82 36 11
18	16691	WV	WA	CE	CA	-	62924930	38 23 00	82 34 11

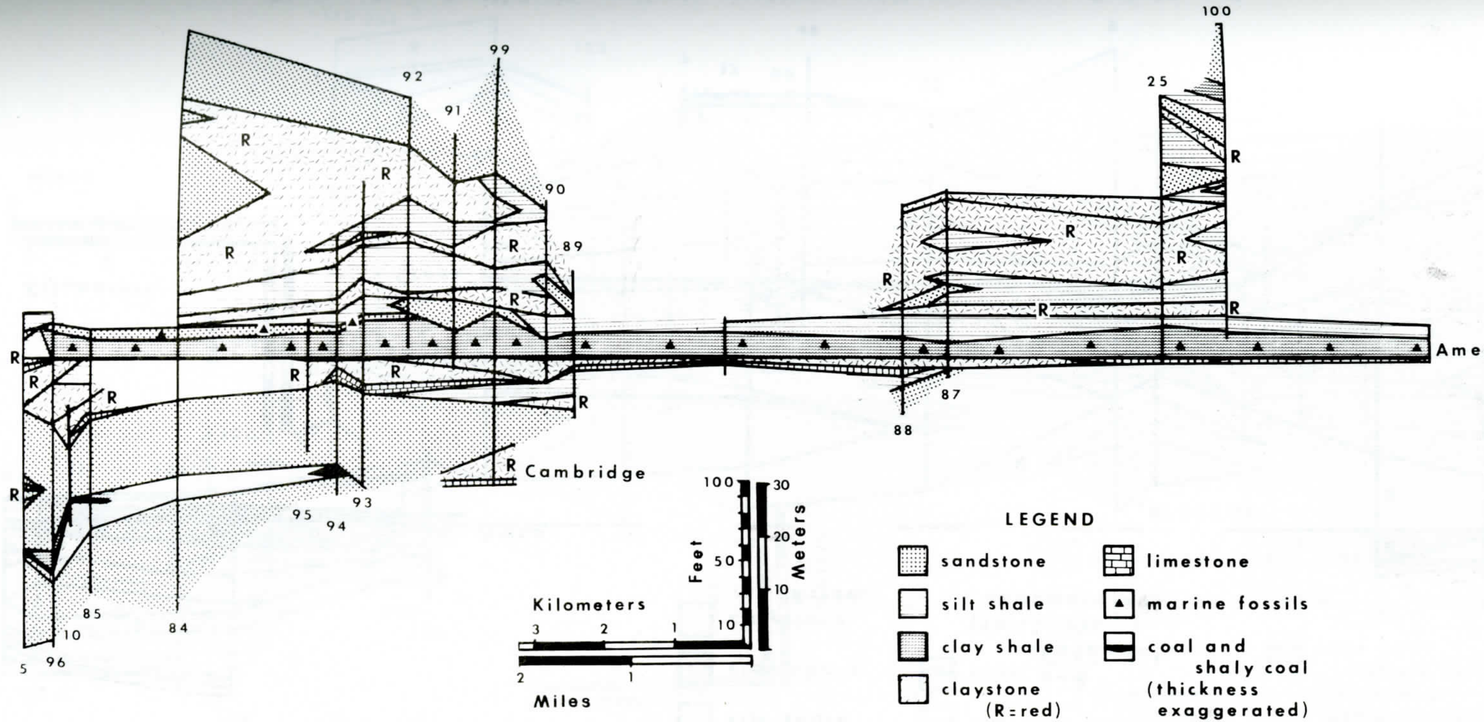


Figure 3. East-West geologic section. This diagram employs modified standard lithologic symbols and uses the base of the Ames Member as its datum. It is constructed in the conventional manner with horizontal distances taken directly from locality to locality as shown in Figure 1.

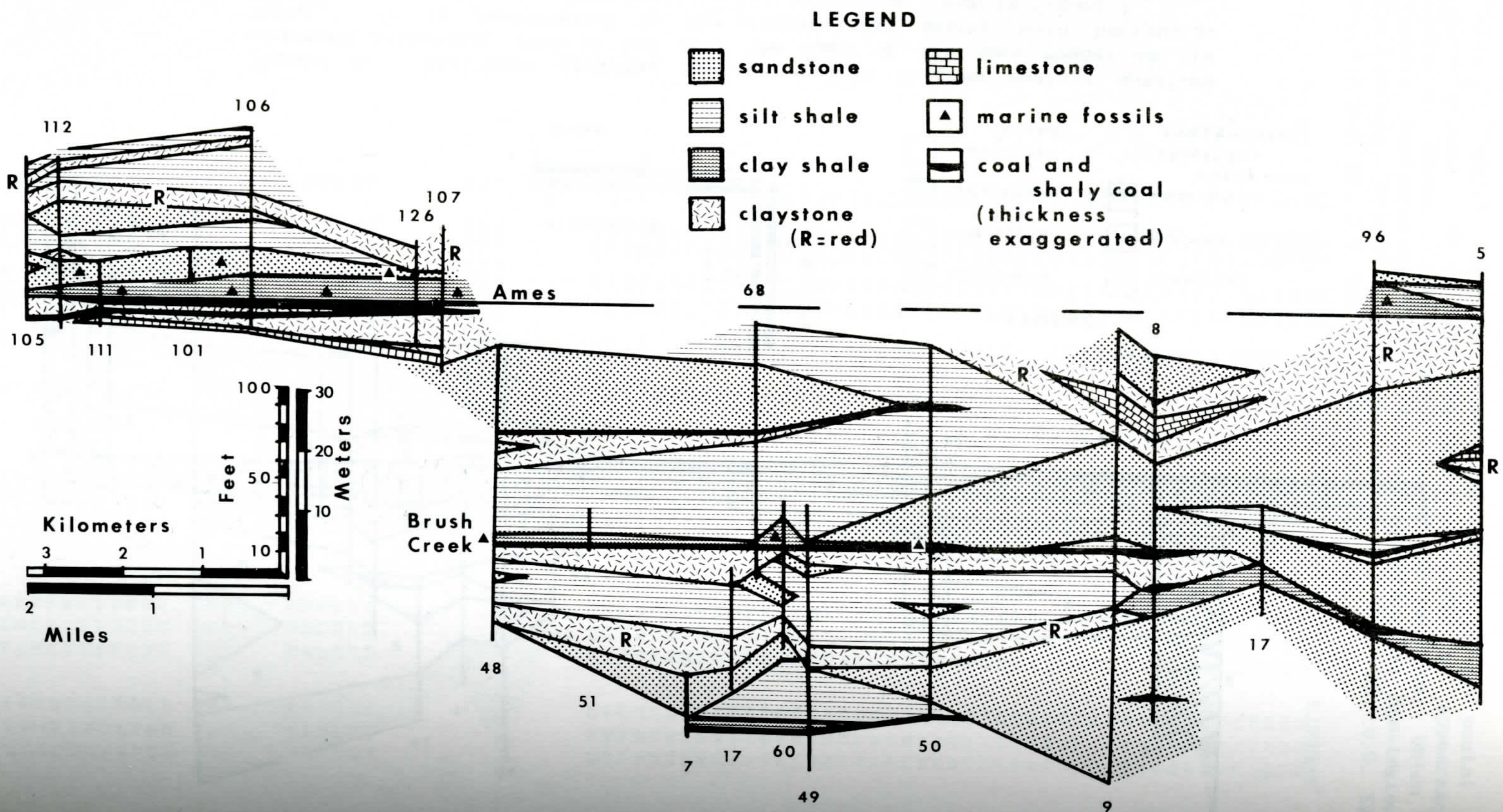


Figure 4. North-South geologic section, northern part. Symbols and correlation datum are the same as for Figure 3. To avoid excessive lengths engendered by zig-zagging across the Big Sandy River numerous times, the line of the section lies midway between the two lines of measured sections on opposite banks and each locality is connected to this line by a perpendicular.

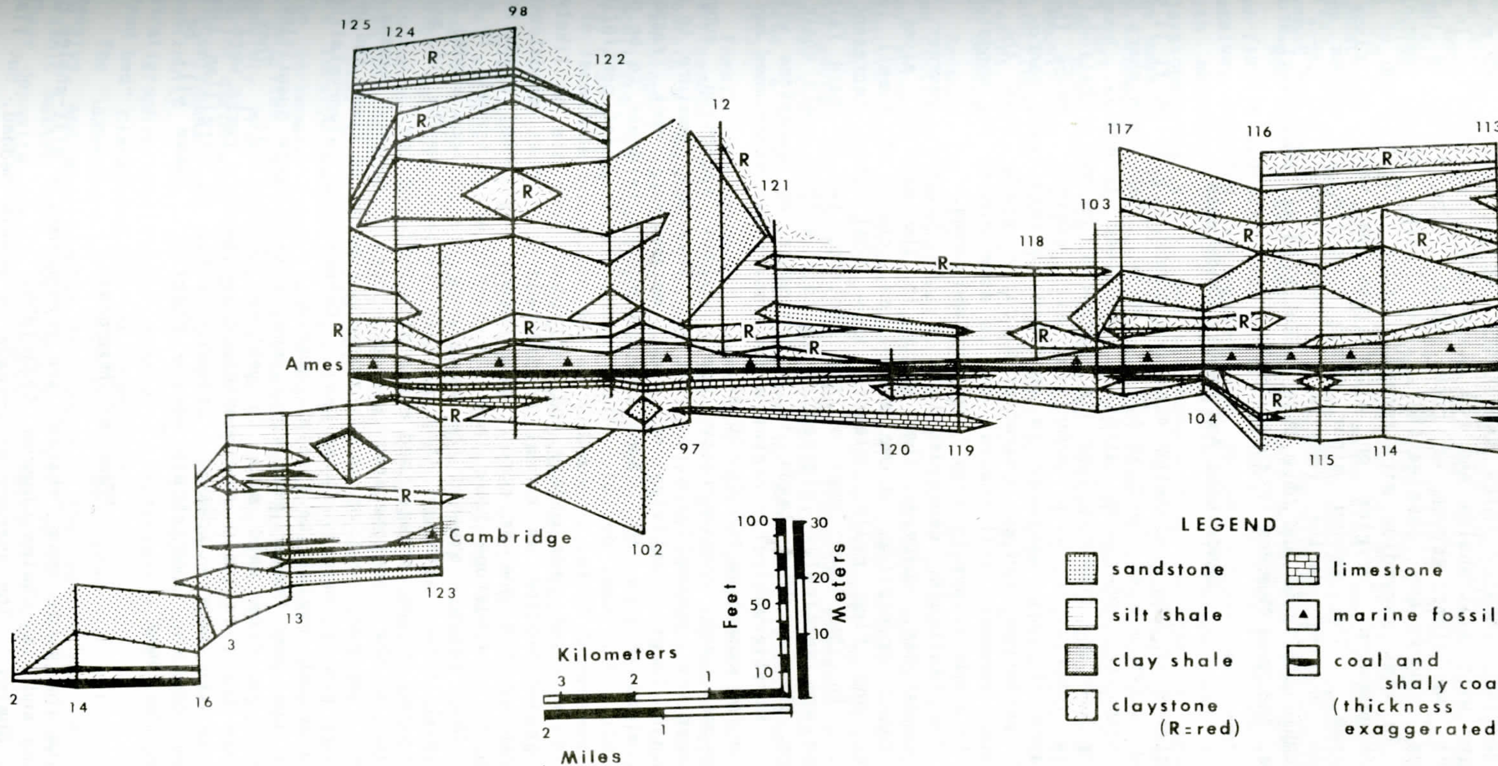


Figure 5. North-South geologic section, southern part. This section is constructed in the same manner as Figure 4.

in the section, locally thick sandstones, up to 100 ft (30 m) or more, interfinger with silt-shales and red claystones ("red beds"). In addition, there are two other groups of rock types. The first is the nodular limestones, underclays (seatearths), and coals, which commonly occur together in that ascending stratigraphic order. This group of rocks has chemical/biochemical origins or alterations as well as close genetic associations. The second minor group consists of shales and impure carbonates bearing marine fossils. These occur at three distinct stratigraphic positions and have been indentified as (ascending) Brush Creek, Cambridge, and Ames Members (Figures 3-5).

Sandstones And Pebbly Sandstones

Sandstones in the Huntington area are commonly 5 to 30 ft (1.5 to 9 m) thick, but a few are as thick as 100 ft (30 m). Many have sharp bases that clearly truncate underlying strata. Followed laterally, however, the truncating base may commonly lose its sharpness, become gradational, and the sandstones interfinger with underlying silt shales. Hence the erosive significance of these surfaces is confined to relatively narrow channels. Tops of sandstone bodies characteristically grade into the overlying strata, most commonly silt-shales, and in some places the sandstone can be observed to grade laterally into finer-grained rocks.

The lenticular cross-sections of the thicker, coarser, and more massive sandstones, wherever they occur in the section, along with the limited basal truncation, slump structures, casts of fossil logs, coal stringers, and flow rolls, clearly show that they represent fluvial channels. In the lower part of the section, two or more fairly thick sandstones are vertically "stacked" with only a thin coaly shale interval separating them. This leads to a dominance by sandstone within this interval. Stratigraphically higher a few individual sandstones may be still thicker, a few exceeding 100 ft (30 m).

Sandstones and conglomerates are, without exception, subgraywackes (sublitharenites, subphyllarenites, etc.). They are immature to submature and contain large quantities of clay, metamorphic rock fragments, and abundant micas (Figure 6H). Some contain enough feldspar to be designated "feldspathic", but most do not. These rocks are at least partially derived from metamorphosed rocks, presumably to the south and southeast of the study area. Grain sizes vary widely and individual beds are as likely to have reverse graded-bedding as normal graded-bedding. Most sandstones in the study area exhibit one or more fining-upward sequences; gradation at the very top of a sandstone body into the overlying silt-shale is especially common. When freshly exposed the sandstones are various shades of grayish and greenish blue, although they quickly (within months in many cases) weather to red, brown, orange, and especially buff and tan.

Clasts in the conglomerates include locally derived masses of muddy sand, mud, lime mud, peat or poorly preserved molds of stems and branches of terrestrial plants (mostly *Calamites*). Others are polycrystalline (meta) quartzite or holocrystalline ("vein") quartz.

Above the Ames Member some sandstones contain large quantities of chert clasts ranging from sand size to pebbles (Figure 6B). They therefore demonstrate the relatively unmetamorphosed sedimentary rocks were exposed to erosion in the source area(s), probably still in Virgilian time. Some sandstones contain invertebrate marine fossils. These will be discussed under "marine rocks".

"Shales" (Mudrocks)

Within the study area "shales" are predominantly silt-shales with fewer mud-shales and clay-shales (Ingram, 1953;1954).

Silt Shales: The average silt-shale is poorly bedded, but fissility is more evident on weathered surfaces. Mudrocks directly above a coal, whether containing marine fossils, or brackish/fresh water fossils, or none, tend to be more fissile than those below the coal. The rocks above coals also contain the largest concentrations of organic carbon of any of the mudrocks studied, and are various shades of gray. Both the unweathered and weathered colors of the silt-shales are similiar to the genetically/provenance-related sandstones with which they are associated; mostly blues and grays when fresh, and olives, buffs, and tans upon weathering. Sandstone interbeds are common and suggest lateral interfingering with sandstone bodies.

"Red Bed" claystones: Rocks here designated "red bed claystones" are non-fissile mudrocks made up predominatly of clay-sized particles. They are massive, unbedded, variegated, red, maroon, purple, and mustard yellow rocks with abundant nodules and pellets of limestone and hematite. Some claystones are silty, and all have a fracture that is vaguely conchoidal. The massive unbedded character of the "red beds" quite likely results from soil-forming processes. Carbonized remains of plants are not preserved, due in part to the highly oxidizing nature of the sediments, but some root casts have been observed. Boundaries between "red beds" and other mudrocks commonly are gradational both vertically and laterally, although distinct interfingering with little gradation is also present. Where gradation is present, it generally is manifest by change in grain size as well as color and bedding.

Because the red color occurs at all depths of burial, including cores and cuttings from deep wells, it is clearly primary or early diagenetic, and may actually fade when exposed to prolonged weathering on the outcrop. Hematite and limestone pellets and nodules probably represent secondary concentrations of mineral matter by such processes as calichification or the biochemical action of plant roots. Subaerial exposure or oxidation under vadose conditions is probably responsible for the bright colors of these rocks, a conclusion also reached by Beerbower (1961) and Duff and Walton (1973).

Fossils are generally rare and poorly preserved in all the non-marine terrigenous rocks. Some of the finer grained and more organic-rich shales contain a compression flora of the leaves of various ferns and seed ferns. Branchiopods (estherids) have been found at not fewer than a half dozen localities, in some places occuring with the leaves of terrestrial plants in gray clay-shales.

Limestones-Underclays-Coals

Like most of the other rocks in the study area, the limestone-underclay-coal successions show evidence of non-marine origin. The limestones are usually light gray when fresh, have little lateral persistence, commonly wedge out within a single outcrop, and contain relatively few fossils (including ostracods, *Spirorbis*, occasional clams and snails, and rarely fish and amphibian bones and teeth) in addition to probable algal traces. The total biota is compatible with a non-marine origin. These carbonates are locally dolomitic, rarely sideritic, and usually quite argillaceous. The carbonate types are primarily micrites and dismicrites with sparse biomicrites, intramicrites (burrow disruption?), and rarely intrasparites, possibly from the same disruptive source ("autobrecciation"). These non-marine carbonates are widespread, locally thick (<10 ft or 3 m), but very discontinuous and not especially significant in either volume or lateral continuity. They do occur in a large number of sections and at various stratigraphic positions, generally integrading and interbedding with the "red bed" claystones and the underclays.

The underclays are generally light gray in color, little stained, massive, and semi-plastic to plastic. They may be more deeply colored with organic matter than the limestones with which they are commonly associated. These deposits appear to represent shales or claystones extensively altered by the actions of plants growing upon them (alumina enrichment, silica depletion). Fossil content (root casts and *Stigmaria*) and proximity to the superjacent autochthonous coals support this interpretation.

Coals in the Conemaugh are mostly thin and of poor quality. The most significant coal is beneath the Ames Member (Ames Coal of Sturgeon, 1957). Where present (at perhaps half of the Ames localities), this coal is as much as 1.5 ft (45 cm) thick and may be bright and blocky, especially where it is thicker. Where thinner it is generally duller and shalier. A similar coal is present immediately below the Brush Creek Member at some localities.

Marine Rocks

Rocks with marine fossils are rare, thin, but have substantial continuity throughout the area. The three named units are, in ascending order, the Brush Creek, Cambridge, and Ames Members (Figures 3-5). Each

Figure 6 - Oriented thin sections of rocks from the Huntington area.

A - Sparse mixed biomicrite, sample 102ALSC, Ames locality 102, 130X, crossed polars. The fossil in the center is a calcitornellid (pseudopthalmid) foraminifer and ostracods are also present. Irregular areas of iron stain are present throughout the slide, indicating substantial siderite content.

B - Subgraywacke sandstone, 70+ ft (21+ m) above the Ames Member at Ames locality 122, 26X, crossed polars. The grain just left of center is chert; micas and other rock fragments are common, most of the quartz is angular and shows only moderate sorting.

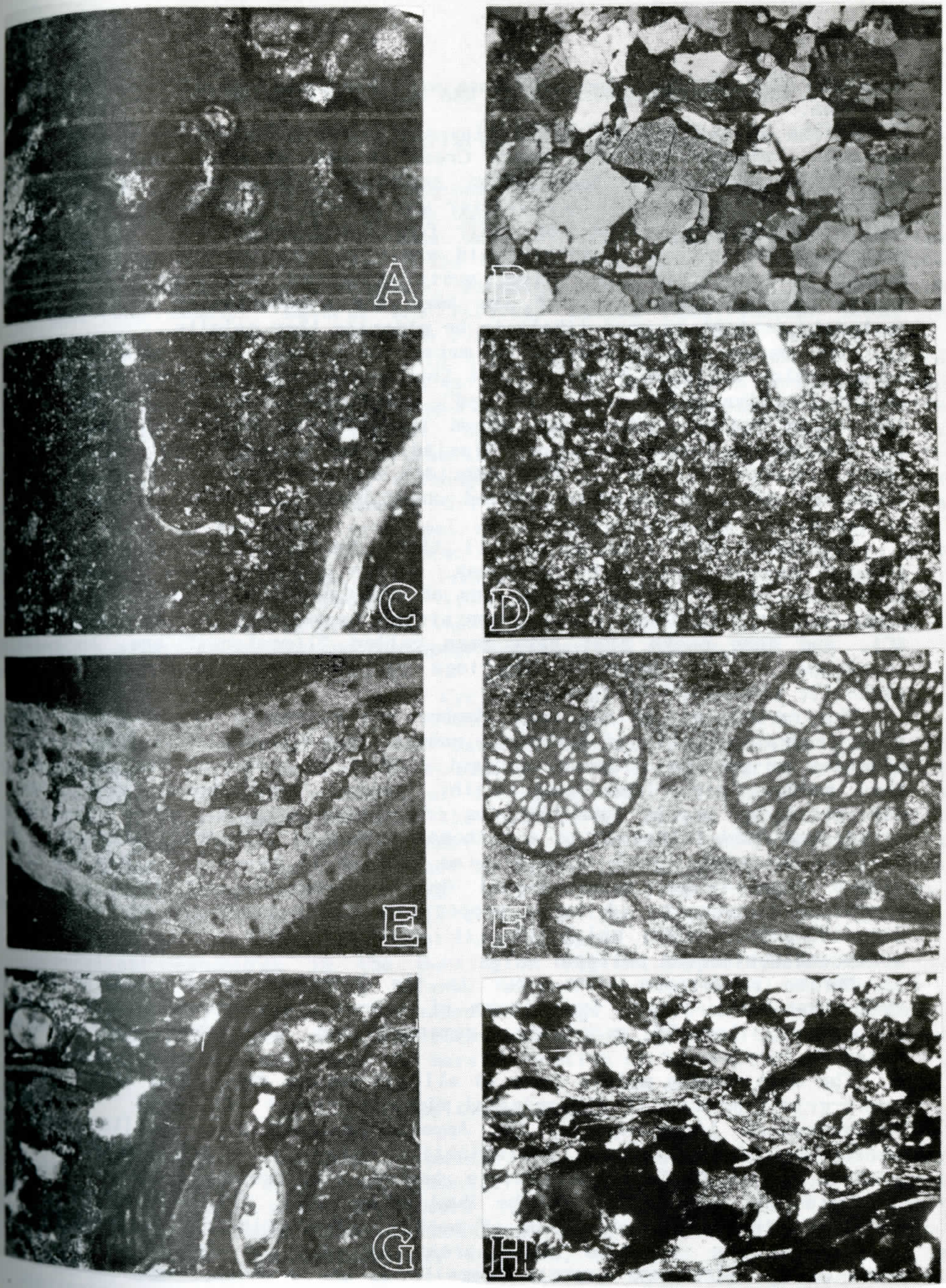
C - Sparse mixed biomicrite/biomicrosparite, sample 97 ALSE, Ames locality 97, 65X, plane light. The matrix consists of lumpy micrite mixed with microspar and numerous iron-stained "pellets" of apparent siderite. Fossils include brachiopods (dominant, as on the left), crinoids, ostracods, and a probable trilobite fragment.

D - Arenaceous siderite, "key bed" just above the Ames at Ames locality 122, 65X, plane light. The rock consists largely of subhedral siderite crystals with large amounts of clay and subordinate sand.

E - Sparse chonetid biomicrite/biomicrosparite, sample 98ALSG, Ames locality 98, 26X, plane light. Other areas of the slide (not shown) are bioturbated, highly sideritic biomicrosparite. In addition to the chonetid brachiopods, ostracods, crinoids, forams, and fish remains are also present in the thin section. Some fossils show micrite envelopes. The brachiopod shown here illustrates three generations of spar infilling; 1) fine fringing spar, followed by, 2) coarse blocky fringing spar coated by a layer of iron stain (while air-filled?), followed by, 3) terminal extremely coarse subhedral spar.

F - Packed fusulinid biomicrosparite, sample 67CLSD, Cambridge locality 67, 26X crossed polars. This rock contains areas that are packed with allochems (fusulinids, crinoids, other fossils) and others that are sparse. Substantial iron stain is present associated with coarser areas of microspar, apparently bioturbated. The entire rock is sideritic.

G - Packed mixed biomicrite, sample 98ALSG, Ames locality 98, 26X crossed polars. This is the same thin section as E (above). Ostracods (the one at the right containing a geopetal), mollusks



(the one near the left edge has a micrite envelope), and a large fish fragment (spine?) are evident in this field.

H - Subgraywacke sandstone, near the base of Ames locality 96, 26X, crossed polars. There are abundant mica grains along with fairly well sorted quartz grains and some feldspar (not visible in this field), both microcline and plagioclase.

consists of both shale and carbonate with sandstone present in the Cambridge and Ames.

Marine shales are generally clay-shales to mud-shales with a few silt-shales. Shales in the Brush Creek Member tend to be organic-rich, especially near their bases, and somewhat calcareous. Shales in the Cambridge marine interval are siltier and generally resemble most of the non-marine silt-shales except for their fossil content. They are somewhat to highly sideritic. A green chonetid shale at or near the base of the Ames Member is relatively less silty, quite calareous, and locally carbon-rich. Shales in the upper part of the Ames are substantially siltier and any carbonate content is more likely to be siderite than calcite.

Some marine shales are red or maroon in color. Their primary brilliant maroon color and good fissility set them apart from the other shales as well as the massive non-marine "red bed" claystones. Similar red shales also containing marine fossils are known lower in the section, especially from the Pennington Formation (Upper Mississippian) a few kilometres to the west (Ferm and others, 1971). The Cambridge and Ames red shales are thought to represent oxidized sediments formed under conditions of subaerial exposure, as lagoonal/tidal flat deposits.

Carbonates are present in all three marine units, although only as nodular masses in the Brush Creek. Thin and lenticular limestones and ironstones form significant portions of the Cambridge and Ames Members. All the limestones in these members contain at least some iron carbonate (Figure 6C), and some rocks that have been called "limestones" are, in reality, ironstones. All observed Cambridge carbonates are siderites or highly sideritic limestones (Figure 6F).

The carbonates in the Ames Member are extremely fossiliferous (Figure 6G), generally highly bioturbated, nodular and discontinuous limestones with moderate or greater siderite content. Most of those in the lower part of the member are limestones with limited persistence, and nodular, discontinuous, poorly fossiliferous ironstones occur higher. The highest widespread Ames lithosome is a carbonate-cemented crinoidal sandstone. This unit is extremely thick locally, (as much as 18 ft, 5.5 m), and the cross-bedded crinoid and other fossil debris bring the carbonate content to roughly 50% overall. A large proportion of the carbonate is siderite, and due to differential solubility it weathers rusty and rounded, with a honeycomb weathering pattern.

Marine sandstones also occur in the Cambridge Member. Thin sections from that unit differ from those of non-marine sandstones only in being carbonate-, largely siderite-, cemented and that some show prominent crinoids.

The most common matrix type in all the marine carbonates is microspar, a grain-growth product of the original micrite (Figure 6E). Most Pennsylvanian limestones in the Appalachians are microsparites, probably reflecting fresh water contamination syn- or epigenetically (Folk, 1965:41,42). Although sparse to packed biomicroparites and biomicrites dominate the limestones in the Huntington area, the most southerly Ames locality (125) contains a lens of poorly washed crinoidal biosparite. Most limestone thin sections show areas of micrite or microspar matrix surrounding allochems, normally fossils of one or more kinds, commonly with regions of stained, siderite rhomb-bearing microspar as concentrations in burrows (Figure 6A). In the southern part of the area a prominent, nodular, sideritic "key bed" occurs a few centimetres above the highest occurrence of marine fossils in the Ames Member. It can be traced for several kilometres as a rounded-weathering, blood-red sandy bed that is dark bluish green on freshly exposed faces (Figure 6D). No marine body fossils have been found in this bed, although it shows evidence of bioturbation.

LITHOGENETIC MODELS AND DEPOSITIONAL ENVIRONMENTS

Lateral variation of rock types

None of the rock types, whether the dominant and thick silt-shales and sandstones or "red beds" on one hand, or the thin coal-underclay-non-marine limestone, or the marine beds on the other, are completely continuous throughout the study area. The marine beds have the greatest lateral persistence and the silt-shale, sandstone, and "red bed" claystone units have the least. Lateral and vertical facies relationships among these complex lithosomes are shown in Figures 3-5. Sandstones tens of feet (or metres) thick "pinch" out within a kilometre (mile) or two. Individual sandstones more than 20 ft. (6.4m) thick can be seen to terminate within a single roadcut in distances measured in tens to hundreds of metres (feet). "Red beds" are not persistent, and generally grade laterally into silt-shales within comparable horizontal distances. The silt-shales, in turn, cannot be traced very far, although they have somewhat greater persistence than "red beds" or sandstones. Silt-shales show rapid lateral facies change into each of the other principal rock types and these intergradations and interdigitations make their tracing difficult. The coal-underclay-non-marine limestone package shows limited lateral persistence both as a group and as individual lithosomes. It appears that the limestone is likely to have the greatest extent and the coal the least, although situations in which all three appear and disappear together are common.

Cyclothems

Despite the low level of lateral continuity shown in these rocks, Beerbower (1961), working with similar rocks in a nearby area, assigned individual recurrent packages of lithosomes to "cyclothems". In a later study in the same area with the benefit of better exposures, Beerbower (1969) demonstrated that the individual lithosomes, and indeed the packages he called "cyclothems", could not be laterally contiguous with the rocks in those area where the "cyclothems" previously had been named. Although the rocks he studied were distinctly different than those that had received the names originally, Beerbower continued to apply the "cyclothem" names. The excellent exposures in the Huntington region support Beerbower's later conclusion about limited lateral extent of individual lithosomes and packages of lithosomes. Increased knowledge about the vertical repetition and lateral changes in these rocks demands a more suitable environmental model or set of models than the mere invocation of "cyclicity".

Depositional Models

Environmental/lithogenetic models for the deposition of Carboniferous rocks in several areas have gradually replaced the formerly widely accepted cyclothem model used over much of North America. In the central Appalachians, a model that has been widely employed was proposed by Ferm and Cavaroc (1969) and subsequently enlarged, detailed and extended in later works, for example, Ferm and others, (1971), Horne and others (1978), and the compendium by Ferm and Horne (1979). In essence this model proposes that the Mississippian rocks, and particularly those a few kilometres west of the present study area, consist of prodeltaic siltstones, "offshore" carbonates, oolite shoals, and some back-shoal lagoonal/tidal flat deposits. Basal Pennsylvanian rocks are quartzose barriers and their back-barrier lagoons, constituting the wave-reworked portions of terrigenous materials transported into the area by prograding deltas from the south and southeast. Through time, the entire suite of environments shifted northward as the clastic wedges built forward from the progressively rising Appalachian Mountains, whose formation was produced, it is generally believed, by the collision of the North American Plate with Europe and/or Africa.

Most of the stratigraphically higher Pennsylvanian rocks studied by Ferm and his colleagues and students in eastern Kentucky consist of deltaic deposits representing primarily lower delta plain to the west (and older), transitional to upper delta plain eastward and upward (Ferm and others, 1979b). These rocks directly underlie those of the present study area and consist of the normal deltaic package of coals, underclays, shales, sandstones, and thin marine intercalations (Ferm, 1970; Ferm and others, 1979b).

Depositional Environments Of The Conemaugh

The Conemaugh rocks in the Huntington area mark two major departures from the general lithologies of these older rocks. First, there is the increased volume of sandstone near the base of the Conemaugh and, second is the introduction of "red beds" somewhat above the thick sandstones.

With thicknesses locally of more than 100 ft (30 m) of complexly truncating sand bodies, interspersed with only thin coals and shales, dominance by sandstone in the lower part of Conemaugh section is nearly total. These lower sandstones mark what could be called a transitional set between "upper delta plain facies" below, and the "lower alluvial plain facies" above. The vertical "stacking" and relatively narrow lateral extent of such thick, massive sandstones indicate repeated and relatively persistent occcupation of the area by large streams depositing cross-bedded sands on their point bars during lateral migration.

Rocks above the lower Conemaugh sandstones have some characteristics similiar to those below, but some are distinctly different. Sandstones higher in the section are clearly of fluvial origin and are commonly quite thick, but there are fewer of them and they are more widely spaced laterally than those in the lower part of the section. These fewer channels represent river conditions where meandering was confined to a more narrow pathway for a protracted length of time. In other respects these sandstones exhibit characteristics consistent with riverine processes: point bars, swales, and repeated lag concentrates of pebbles and driftwood.

Blue-gray silt shales represent overbank deposits and some can be seen to diminish from silt-shale to mud-shale to clay-shale away from a prominent channel. Some of these can also be observed lapping up on the flanks of large sand bodies where they clearly represent natural levee deposits. The finest grained rocks represent margins of the floodplain only rarely inundated.

The primary points of difference between the middle-upper Conemaugh rocks and the upper delta plain facies beneath are the presence of the "red bed" claystones, the absence of well developed coal beds and the presence of a widespread marine bed, the Ames Member. The presence of the "red beds" commonly has been explained by recourse to a new source for the sediments or some dramatic change in climate (see references in Turner, 1980, for examples). Indeed, recent opinion supports increasing aridity during this part of the Carboniferous (Busch and Rollins, 1983; Phillips and others, 1983; Sigleo, 1983; and Schutter and Heckel, 1983). Although it is true that a climatic change could produce red coloration in the fine-grained detritus, as well as thinner, more poorly developed coals from changes in the vegetation, it is equally likely that the greater degree of oxidation reflected by these attributes could result from better drainage of the alluviated surface. Instead of deposition at or below sea level under totally waterlogged conditions, middle and upper Conemaugh rocks apparently were formed on flood plains distinctly above sea level. The variegated claystones were deposited by flood waters at substantial distances from the stream channels and represent the finest sediments carried beyond the natural levees where the silt-shales were accumulating. Under these conditions on a vadose alluvial plain, high levels of oxidation could only lead to primary red-mottled coloration in contrast to the phreatic blue and gray sediments accumulating downstream on the continuously saturated delta. Repeatedly and intermittently, the flood plain clays were soaked and dessicated. Wetting was largely by slackwater flooding and dessication developed from high levels of evaporation and produced calichification (limestone pellets and nodules). Beerbower (1961) reached a similar conclusion about the "red beds" he studied higher in the section and somewhat east of Huntington.

SUMMARY

Rocks exposed in the immediate vicinity of Huntington belong to the Conemaugh Group and slightly subjacent strata and represent upper delta plain, transition, and oxidized lower alluvial plain deposits. Thus these rocks, and the substantially similar overlying Monongahela and Dunkard Groups, constitute the final chapter, upstream and upsection, for the progradational Appalachian Carboniferous model proposed by Ferm and others, (1971). Individual beds have little persistence, and lithologies recur upward in an almost monotonously repetitive (rather than cyclic) manner. The most important non-marine rock types encountered and their principal lithogenetic origins (parentheses) are: silt-shale (overbank deposits), sandstones (channels), "red bed" claystones (dominantly vadose floodplains), underclay or seatearths (leached soils), limestones (lacustrine), and coal (paludal).

Marine rocks constitute a volumetrically minor part of the total section and are generally thin. The two older marine units, the Brush Creek and Cambridge Members, are thin (<2 m) and are not geographically widespread. The youngest, the Ames Member, is thicker and is present virtually throughout the study area. The Brush Creek and Cambridge Members were deposited following trangressions into upper delta plain or transitional environments. Unlike all previous marine episodes, the Ames transgression inundated well-established fluvial conditions.

Although climatic changes during the Late Carboniferous may have led to drier, more arid climates, the progradation of major clastic wedges led to more upstream environments replacing more downstream ones through time. Thus better drained, oxidized, vadose alluvial plain conditions replaced waterlogged, reduced, deltaic ones and may partially explain the apparent increase in aridity.

ACKNOWLEDGEMENTS

During this long-term project, a great many people made real and appreciated contributions. Acknowledgements are alphabetically listed. Susan A. Bridges did a senior thesis dealing with Ames correlations at the College of Charleston in 1981; a Lilly Grant from the College of Charleston provided support facilities, office space, and funding for drafting; Prof. and Mrs. George B. Corrie allowed me to use their home as a base of operations both in 1965 and 1979; John C. Ferm supervised the initial project and continued to provide inspiration and support during the intervening years; Teresa C. Garrett drafted and organized several aspects of the study; Michael C. Hansen, Ohio Division of Geological Survey, served as a point of contact with his organization, provided stratigraphic information, and read and commented on the manuscript; Ethel M. Jackson, University of Houston-Downtown, produced the final several versions of the typescript; James E. Jordon provided many gratefully appreciated incidental services during the writing phase of the study; (now Dr.) D. T. Long did some of the initial correlation work on the 1965 sections as a part of an undergraduate project at Monmouth College in 1969-70; Joan Newell typed the measured sections of OGS forms; Suzanne Truesdale provided computerized strip logs.

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THE STRATIGRAPHY, PALEOGEOGRAPHY, DEPOSITIONAL ENVIRONMENT, FAUNAL COMMUNITIES, AND GENERAL PETROLOGY OF THE MINNEHAHA SPRINGS MEMBER OF THE SCHERR FORMATION, AN UPPER DEVONIAN TURBIDITE SEQUENCE, CENTRAL APPALACHIANS.

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ABSTRACT

The Minnehaha Springs Member of the Upper Devonian Scherr Formation is here named and described. This clastic bundle of micaceous coarse-grained siltstones can be identified in both outcrop and subsurface for 235 kilometers (150 miles) along the Allegheny Front from southern Pennsylvania to east-central West Virginia. The Minnehaha Springs Member is 6.37 to 30.0 meters (20.9 to 98.4 feet) thick, consisting of interbedded very thinly to thickly bedded, medium gray siltstones and olive gray shales, with some grayish-red siltstones and shales.

The Minnehaha Springs Member was deposited by turbidity flows situated in lower middle fan environments. These turbidity flows originated from the east from three major, persistent depositional systems: the Fulton, Mouth of Seneca, and Augusta. Three faunal communities are associated with the submarine fan environments of the Minnehaha Springs Member. The Burrowing Community is a pre-turbidite fauna associated with outer fan and basin plain environments. The Pelmatozoan-*Ambocoelia-Chonetes* Community may have colonized outer fan and lower middle fan environments immediately after turbidite flows. The *Loxonema* Community represents a transported thanatocoenosis localized near the discharge of channels in the lower middle fan, or suprafan environments.

The Minnehaha Springs Member may be used as a time band within the marine strata of the Devonian Catskill Delta Complex of the Central Appalachians. The Minnehaha Springs Member may be the shoreward equivalent of the gas productive Sycamore siltstone found in north-central West Virginia.

INTRODUCTION

A shaly grayish "redbed" zone and an associated coarser-grained clastic bundle near the base of the Cohocton Stage was first recognized by Dennison (1970) based on field work done principally in 1962 and 1963. This "redbed" zone was used as the contact between the Upper Devonian Brallier and Scherr Formations north of LaVale, Maryland. Geologic mapping in south-central Pennsylvania by de Witt (1974) placed these "reds" in the top of the Brallier Formation, whereas Dennison (1971) placed them in the lowermost Scherr Formation. In either case, there are no "redbeds" older than these in the Devonian sequence of the study area, and the next younger redbeds occur some 366 meters (1200 feet) above the Minnehaha Springs Member in the lower portion of the Mallow Member of the Foreknobs Formation (Dennison, 1970, Figure 3). Kulander (1968, p. 37) recognized the existence of brick-red shales in the middle of the Brallier Formation exposed in the Browns Mountain anticlinorium of West Virginia and used them as a stratigraphic marker in geologic mapping.

The study area of this report (Figure 1) extends south from Bedford County, Pennsylvania to Greenbrier County, West Virginia, a distance of approximately 235 kilometers (150 miles). Many of the measured sections are located along the Allegheny Front. Others are located in the Browns Mountain anticlinorium and in the Bedford, Clearville, and Middle Mountain synclines in West Virginia. A total of 27 sections were measured, all

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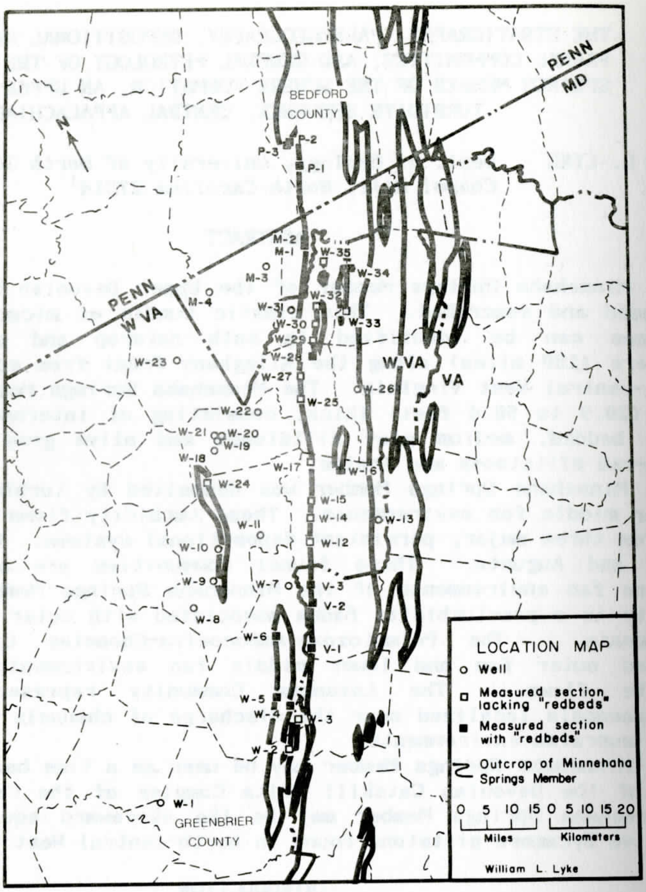

Figure 1. Location map of the study area. Modified slightly from Lyke, 1981 (Figure 1).

containing the coarser-grained clastic bundle and/or the associated grayish "redbeds". Data from 17 wells supplied both isopach data for the Minnehaha Springs Member and gamma ray signatures for correlations east and west of the outcrop area. Appendices containing the geographic location and the data obtained from the measured sections and wells have been presented in Lyke (1981). The terms "very thinly", "medium", and "thickly" bedded used in this report will follow the definitions proposed by Ingram (1954).

The Fulton and Augusta depositional systems recognized in this report were originally named the Fulton Lobe (Willard, 1934) and August Lobe (Dennison, 1970). Grant Bay, which lies between the Fulton and Augusta lobes, was originally named by Dennison (1971) and includes the Mouth of Seneca depositional system. The Mouth of Seneca Lobe was first recognized by Kirchgessner (1973).

"Depositional system" or "system", is used to describe the three major source areas of sediment in this study area as these systems contain one or more depositional lobes, suprafan lobes, or suprafans. Measured sections which represent the Augusta depositional system extend from the Minnehaha Springs (W-4) to Route 250 (V-2) sections. Sections from Ketterman Knob (W-15) to Maysville (W-25) represent Grant Bay, and sections from Ridgeville (W-30) to Glade Pike (P-3) represent the Fulton depositional system.

STRATIGRAPHY AND DESCRIPTION OF UNITS WITHIN STUDY AREA

Brallier Formation

The thickness of the Brallier Formation (Butts, 1918) is reported (Dennison, 1970) to be from 175 to 662 meters (from 575 to 2170 feet) along the Allegheny Front from U.S. Route 250, Virginia to Corriganville, Maryland. The shortening of the sections along the Allegheny Front is primarily due to faulting (Dennison, 1971; Berger et al, 1979). The Brallier Formation consists of interbedded siltstones and thickly laminated shales. Siltstone thickness in the Brallier range from 3 centimeters (0.1 foot) to approximately 30 centimeters (1 foot). They are characteristically light gray (N7) to medium gray (N5) fresh and tend to weather moderate yellowish brown (10YR5/4).

The Minnehaha Springs Member is stratigraphically the lowest major coarser-clastic bundle in both the Brallier and Scherr Formations other than the Back Creek Siltstone (Avary, 1978; Avary and Dennison, 1980). The Back Creek Siltstone is present only in the central and southern parts of the field area and lies approximately 100 meters (328 feet) below the Minnehaha Springs Member (Figure 2). Complete or nearly complete stratigraphic

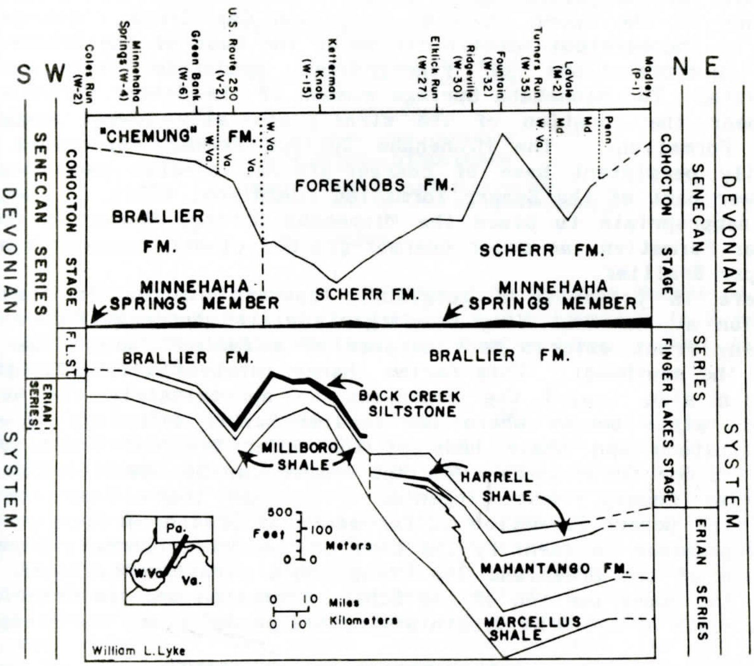

Figure 2. Stratigraphic cross-section of Upper and Middle Devonian strata along and near the Allegheny Front. Both the Scherr and Brallier Formations thin toward the Southwest as a result of stratigraphic convergence. Datum is the base of the Minnehaha Springs Member. Southwest of the Virginia - West Virginia border along the Allegheny Front, the Minnehaha Springs Member lies totally within the Brallier Formation. The Minnehaha Springs Member lies at the base of the Scherr Formation toward the northeast. The diagram is designed after Avary, 1978 (Figure 7).

sections of the Brallier below the Minnehaha Springs Member were encountered at the LaVale (M-2), Zeke Run (W-17), and Minnehaha Springs (W-4) sections. The Brallier has been described by many authors (Dennison, 1971; Glaeser, 1979; Kepferle et al, 1978; Lundegard et al, 1978; and others) as deposited by distal deep water turbidity currents originating from the east.

Scherr Formation

The Scherr Formation is the lowest stratigraphic unit identified in the Upper Devonian Greenland Gap Group (Dennison, 1970). The Scherr Formation ranges in thickness from 139 to 617 meters (456 to 2025 feet) and everywhere appears conformably bounded by the overlying Foreknobs Formation. Although the Scherr Formation consists predominantly of light olive gray weathering siltstones, considerable fine sandstones and shales are also present. The upper portion of the Scherr is lithologically similar to the Brallier, in that it consists predominantly of interbedded siltstones and shale without sandstone. The lower Scherr is coarser grained, containing a small percentage of fine-grained sandstones. The basal Scherr beds contain *Cyrtospirifer chemungensis* (Conrad) and *Ptychopteria (Cornellites) chemungensis* (Conrad) (Dennison, 1970) which place the age of these deposits at the lowermost Cohoctan Stage (Rickard, 1964).

I here propose to redefine the Scherr Formation so its base is the base of the Minnehaha Springs Member. The base of the Scherr Formation was originally defined by Dennison (1970) to include the stratigraphically lowest sandstone of the Upper Devonian Greenland Gap Group. His definition may result in inconsistent identification of the base of the Scherr Formation as field interpretations of fine-grained sandstone vary with different geologists. The Minnehaha Springs Member of the Scherr Formation lies at or very near the location of the stratigraphically lowest sandstone in the Scherr Formation. The Minnehaha Springs Member represents a distinct, laterally persistent base of coarser-grained clastic sequences which mark the lower part of the Scherr Formation (Dennison, 1971). It would therefore seem inappropriate to place the Minnehaha Springs Member in the top of the Brallier Formation, as major coarser-grained clastic sequences are absent in the upper Brallier.

There is presently a geographic cut-off in use of the name Scherr Formation at the West Virginia-Virginia state boundary (Figure 2) along the Allegheny Front outcrop belt because of a facies change loss of sandstone toward the southwest. This facies change involves a lateral, general fining of grain size toward the southwest to approximately the West Virginia-Virginia state border where the sandier Scherr lithology is equivalent to the siltstone and shale beds of the upper Brallier Formation (Dennison, 1970). Since the Minnehaha Springs Member can be identified as far south as Greenbrier County, West Virginia, I propose that it may be possible to extend the Scherr Formation as far south as Greenbrier County. Although it is now possible to identify the base of the Scherr Formation and, therefore, the base of the Greenland Gap Group, more stratigraphic work is needed to define the upper portion of the Scherr Formation and the Greenland Gap Group to the south of the West Virginia-Virginia border along the Alleghany Front.

THE MINNEHAHA SPRINGS MEMBER

The name Minnehaha Springs Member here designates a sequence of siltstones, shales, and some fine-grained sandstones that can be traced from Bedford County, Pennsylvania, to Greenbrier County, West Virginia. The type section (Figure 3) of the Minnehaha Springs Member is located 1.9 kilometers (1.2 miles) south of Minnehaha Springs, Pocahontas County, West Virginia, on the north side of West Virginia Route 39.

The Minnehaha Springs Member is a zone of interbedded siltstones, shales, and some sandstones that is in part coarser-grained than the Brallier Formation. Very thinly to thickly bedded siltstones containing sedimentary structures of turbidite origin, such as flute casts, sole marks, and abrupt sharp bases and gradational tops, are interbedded with very thinly to thickly bedded silty shales and shales. The siltstones are predominantly medium gray

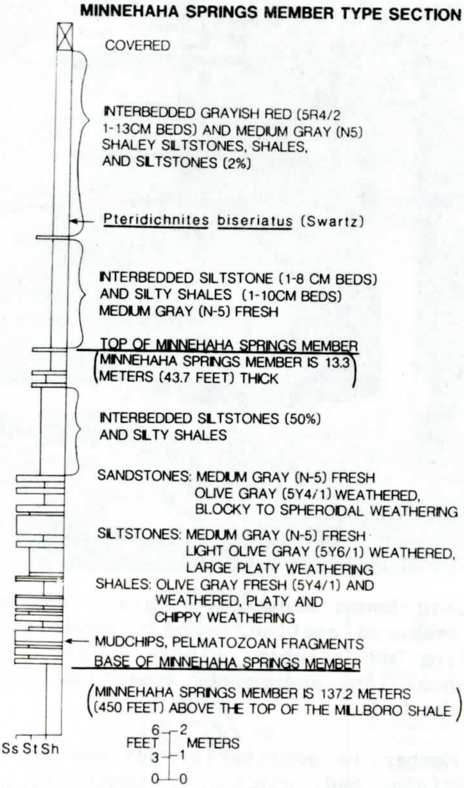

Figure 3. The type section of the Minnehaha Springs Member of the Scherr Formation. The measured section is located 1.9 kilometers (1.2 miles) south of the village of Minnehaha Springs, Pocahontas County, West Virginia, on West Virginia Route 39.

(N5), medium dark gray (N4), and olive gray (5Y4/1) when fresh. This medium gray (N5) to medium dark gray (N4) color is more abundant at Minnehaha Springs (W-4), Green Bank (W06), Brushy Run (W-16), Zeke Run (W-17), LaVale South (M-1), and LaVale (M-2) which appear to be nearer-shore deposits. The shales are olive gray (5Y4/1) to light olive gray (5Y6/1) when fairly fresh. No unweathered samples could be obtained from outcrops. Some grayish red (5R4/2) to very dusky red (10R2/2) beds of sandstone, siltstone, and shale are also present locally. The dominant weathering colors are olive gray (5Y4/1), light olive gray (5Y6/1), moderate yellowish brown (10YR5/4), and grayish red (5R4/2). The shaly intervals between the siltstones and sandstones decrease in grain size from very silty shales in the northeast to predominantly slightly silty shales in the southwest. Although siltstone is the dominant coarser-grained lithology, some sandstone is present in Pocahontas, Pendleton, and Mineral Counties, West Virginia, Allegany County, Maryland, and Bedford County, Pennsylvania.

The Minnehaha Springs Member was deposited by turbidity flows originating from the east. Bouma sequences (Bouma, 1962; Middleton and Hampton, 1973) tend to be incomplete in the Minnehana Springs Member, although a complete Bouma Sequence was observed at the Brushy Run (W-16) measured section (Figure 4).

The top of the Minnehaha Springs Member is defined as the youngest siltstone or sandstone bed equal to, or greater than, 10 centimeters in thickness that lies within 2 meters above a similar siltstone or sandstone bed at least 10 centimeters thick, within the bundle. The base of the

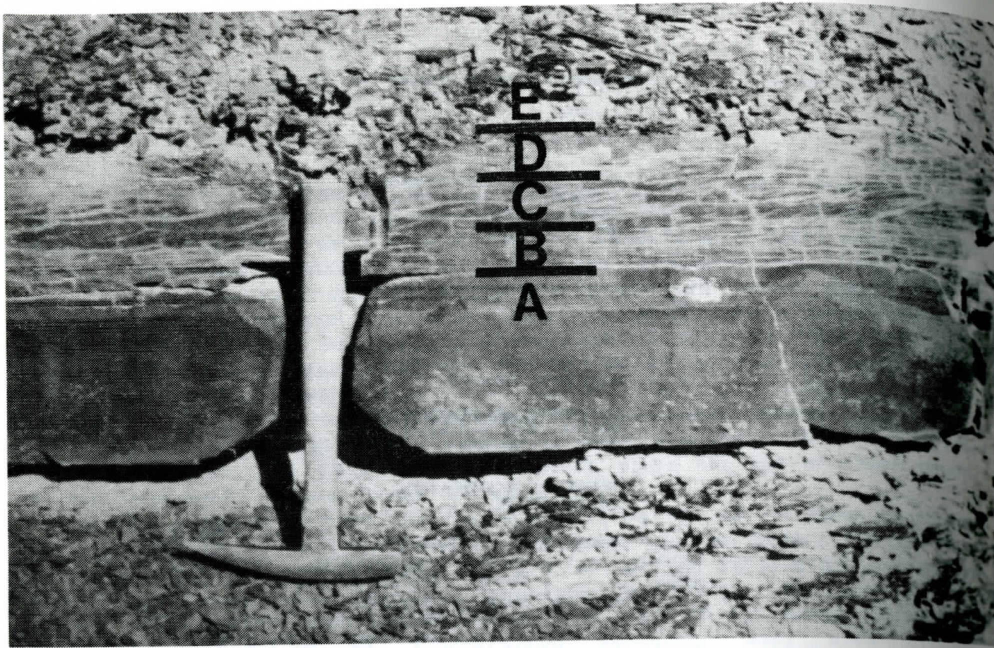

Figure 4. A complete Bouma sequence within a coarse siltstone bed at the Brushy Run (W-16) measured section. The C division of the Bouma Sequence displays cross-bedding and climbing ripple marks. The hammer handle is 28 centimeters (11 inches) long and points upsection. Note the abrupt base and gradational top of the bed.

Minnehaha Springs Member is arbitrarily defined as the base of the oldest siltstone or sandstone bed which is equal to, or greater than, 10 centimeters in thickness, and lies within 2 meters below another 10 centimeter thick siltstone or sandstone bed located within the bundle. Although the 10 centimeter minimum thickness for coarser-grained beds is arbitrary, it conforms to the thinnest bed identified as the Back Creek Siltstone of the Brallier Formation (Avary, 1978). Beds ten centimeters thick were also used by Walker and Mutti (1973) to distinguish between proximal and distal classical turbidite facies. An exception to the boundary definition for the Minnehaha Springs Member occurs in the Grant Bay area. The Minnehaha Springs Member at Zeke Run (W-17), Brushy Run (W-16), Ketterman Knob (W-15), and Briery Gap Run (W-14) contains two coarser-grained bundles separated by three to five meters of predominantly shale. These two bundles were both incorporated into the Minnehaha Springs Member because they are stratigraphically close together and distinct within the large amount of section observed at Zeke Run (W-17).

In outcrop the Minnehaha Springs Member ranges in thickness from 6.4 to 30.0 meters (20.9 to 98.4 feet). The thickest single siltstone or sandstone bed occurs at the type section near Minnehaha Springs. The thickness of the thickest bed for all measured sections ranges from 30 to 88 centimeters (1.0 to 2.9 feet) in the study area (Figure 5). The average (sand + silt)/shale ratio is 1.24 and ranges from 0.56 to 1.67. The base of the Minnehaha Springs Member occurs approximately 776 meters (2545 feet) stratigraphically above the Tully Limestone in the northeast at the Madley (P-1) section. This stratigraphic distance generally decreases towards the southwest to approximately 175 meters (575 feet) at the Minnehaha Springs (W-4) section (Figure 6). This thinning of the stratigraphic interval is apparently the result of deposition of progressively finer-grained sediments toward the southwest, rather than an unconformity.

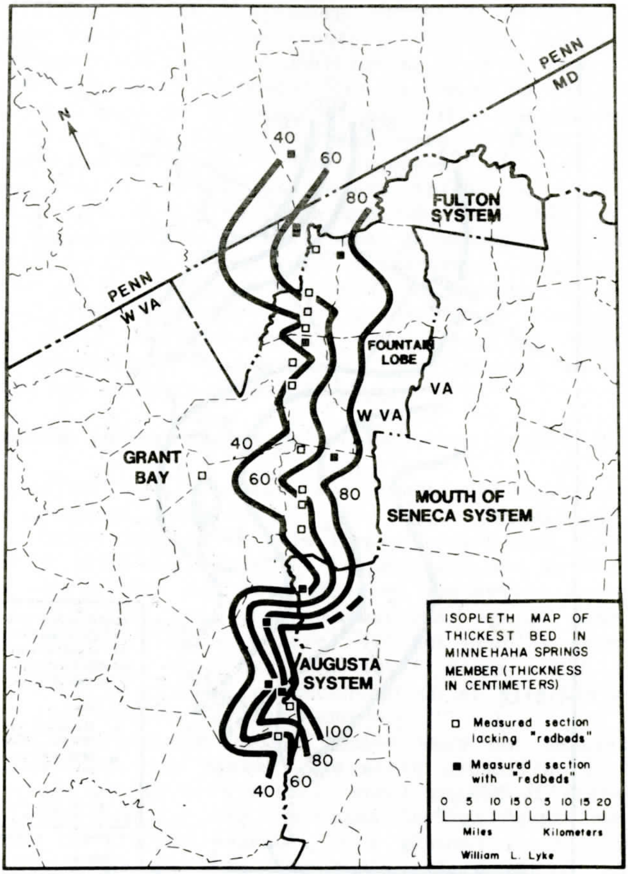

Figure 5. An isopleth map of the thickest siltstone or sandstone bed in the Minnehaha Springs Member. Modified slightly from Lyke, 1981 (Figure 9).

"Redbeds" of the Minnehaha Springs Member

Grayish-red to very dusky-red beds of the Minnehaha Springs Member are locally present in silty shale, siltstone, and sandstone lithologies. The thicker, more resistant, grayish-red turbidite siltstones and sandstones probably represent the original color of the sediment during deposition because 1) these beds are consistently "red" both laterally and throughout the thickness of the bed; 2) the outer weathered zone of these beds are usually grayish to olive gray with the reddish portion in the unweathered center of the bed; 3) the presence of reduced olive green burrows in some siltstone "redbeds" at both the Madley (P-1) and Minnehaha Springs (W-4) sections; and 4) the persistant stratigraphic location of these grayish red siltstones at, or near, the basal portion of the Minnehaha Springs Member in the Huntersville (W-5), Green Bank (W-6), LaVale South (M-1), LaVale (M-2), and Madley (P-1) measured sections. These nearer-shore sediments did not undergo 1) sufficient aqueous dispersion (hydration) during deposition, or 2) total reduction since deposition, therefore retaining their original "red" color.

"Red" silty shales are far more common than "red" siltstones. The "red" silty shales of the Minnehaha Springs Member were deposited as distal portions of turbidity flows. Thickness of "red" silty shale intervals range from 1 to 8 centimeters. These "red" silty shales are located above and/or

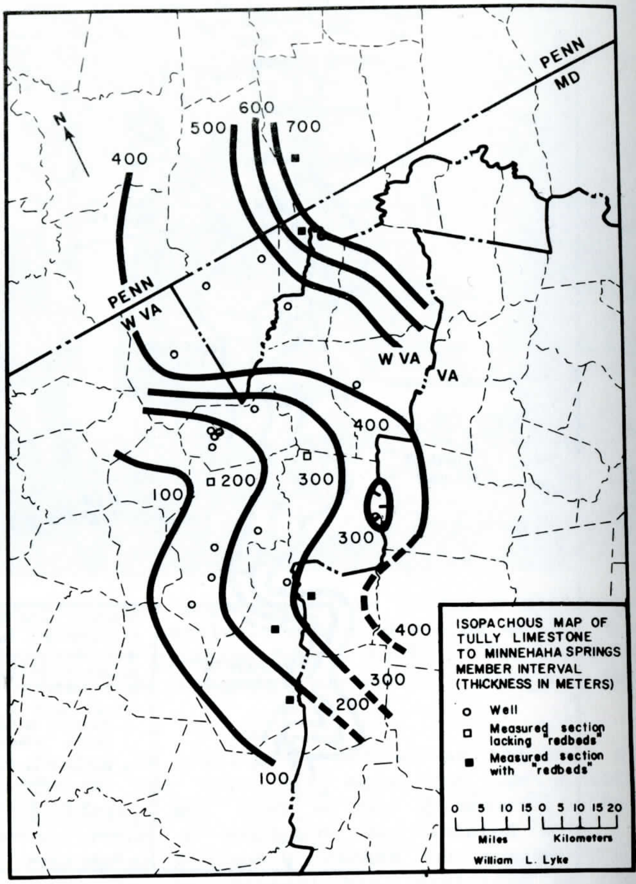

Figure 6. An isopachous map of the stratigraphic interval from the top of the Tully Limestone to the base of the Minnehaha Springs Member. Modified slightly from Lyke, 1981 (Figure 15).

below the Minnehaha Springs Member but rarely within it. Whether the color of these shaly "redbeds" is syndepositional is unclear. The grayish-red color of the silty shales might have occurred as a result of weathering as indicated by the 1) existing gradual color boundaries between the interbedded olive gray and grayish-red silty shales and very thinly bedded siltstones; and 2) the poor horizontal continuity of these "redbeds" across distances of a few meters within a single bed. However, their association with the Minnehaha Springs Member and "red" siltstones in an isolated stratigraphic position implies a syndepositional origin for "redbeds" of all lithologies. Measured sections which contain grayish-red silty shales are located at Minnehaha Springs (W-4), Huntersville (W-5), Green Bank (W-6), Rt. 250 (V-2), Rt. 642 (V-3), Brushy Run (W-16), Knobly Run (W-28), LaVale South (M-1), LaVale (M-2), and Madley (P-1). Grayish-red beds were observed in Bedford County, Pennsylvania at ten locations south of the Glade Pike (P-3) section.

Dennison (1971) and Dennison and de Witt (1972) proposed that these "redbeds" and associated coarser-grained clastic rocks originated as a result of an eustatic drop in sea level. Both the Fulton and Augusta lobes appeared to have a simultaneous development of "redbed" sequences which are separated by hundreds of feet of stratigraphic thickness from other "redbeds". Lowering of sea level would have shifted the shoreline basinward and allowed oxidized reddish near-shore sediments to be deposited farther westward into the basin. Various events of flooding during this sea level drop may have contributed to the deposition of "red" sediment basinward. An example of flood originated "redbeds" can be found in Swift (1976, p. 336-339) who summarized a study by Drake et al (1972) on the dispersal pattern of red-brown silt and clay sized flood sediments on the Santa Barbara-Oxnard shelf off southern California. The thickness of the California flood deposits ranged from 1 to 10 centimeters thick one year after initial deposition. Although direct evidence can not be cited to indicate that similar events are associated with the Minnehaha Springs Member, the similarities such as grain size, color of sediment, bedding thickness, and the areal extent of the deposits in these two studies are striking. If a sudden sea level drop occurred, shifting the shore toward the basin and causing the Minnehaha Springs Member to be deposited, then the dispersal of reworked oxidized "red" sediment by large storms into lower mid-fan environments does not seem unreasonable.

Minnehaha Springs Member as a Chronostratigraphic Unit

The occurrence of these "redbeds" and coarser-grained clastic sequence in a limited stratigraphic position on three major depositional lobes within the Appalachian basin implies a relatively short-lived event, such as an eustatic sea level drop (Dennison, 1971; Dennison and de Witt, 1972), or a large scale regression that may have affected the entire basin. The lack of "redbeds" in the Brallier and Scherr Formations, except near the Minnehaha Springs Member, suggests that sea level did not drop in other parts of the basin during that time, or those sea level drops which did occur were insufficient in size to produce a concentration of slightly coarser-grained rocks similar to the Minehaha Springs Member.

Portions of the Minnehaha Springs Member were not deposited at exactly the same time. Rather, the member represents deposition by three separate depositional systems within a relatively short period of geologic time. The total time of deposition for the Minnehaha Springs Member may have been as brief as 100,000 to 170,000 years. This assumption is based upon the 6 million year time range for the Frasnian (Harland et al, 1964), which is equivalent to the Senecan Series of North America, and the assumption that the Cohocton Stage of the Senecan Series represents 4 million years. The Minnehaha Springs Member lies near the base of the Cohocton Stage and has an observed maximum thickness of 30.0 meters (98.4 feet) at Madley (P-1). The top of the Cohocton Stage is located near the top of the Greenland Gap Group at the base of the Pound Sandstone of the Foreknobs Formation (McGhee and Dennison, 1980). The top of the Greenland Gap Group is 709 to 1185 meters (2325 to 3885 feet) (Dennison, 1971) above the base of the Scherr along the Allegheny Front in West Virginia. Assuming continuous rates of deposition for the Greenland Gap Group and applying ratios, it seems reasonable that the Minnehaha Springs Member was deposited in a time span of approximately 100,000 to 170,000 years, and therefore represents a brief time band. This time band represented by the Minnehaha Springs Member is relatively small and should not affect the use of the member as a time line in large scale correlations.

Identification of a single continuous time line throughout the Minnehaha Springs Member is probably impossible because the member was deposited by three overlapping submarine fans. The absence of a single, discrete key bed throughout the lateral extent of the Minnehaha Springs Member precludes identification of a single, member wide time line. Also, no discrete fossil zone extends over the entire length of the field area.

I feel the best approximation of a time line would be the conspicuous grayish-red siltstone and sandstone beds. The use of these "redbeds" as a time line is questionable as well because they are not continuous through the field area. When present, however, their location at the base of the

Minnehaha Springs Member could represent the initiation of the eustatic, or apparent eustatic, drop in sea level. Therefore, in detailed correlations, the best approximation of a time line is the base of the Minnehaha Springs Member.

PALEOGEOGRAPHY

The paleogeography of the Minnehaha Springs Member is represented by a bathymetrically irregular ocean bottom created by locally higher sedimentation rates on submarine fan depositional lobes. The deposited sediment probably originated initially from three coastal stream systems discharging into the basin from the east. Bathymetric variability during the deposition of the Minnehaha Springs Member was partially controlled by progradation of lower submarine fan deposits over basin plain deposits, and by the interfingering of sediment from adjacent major depositional systems. The generalized geographic position of the three depositional systems is presented in Figure 7.

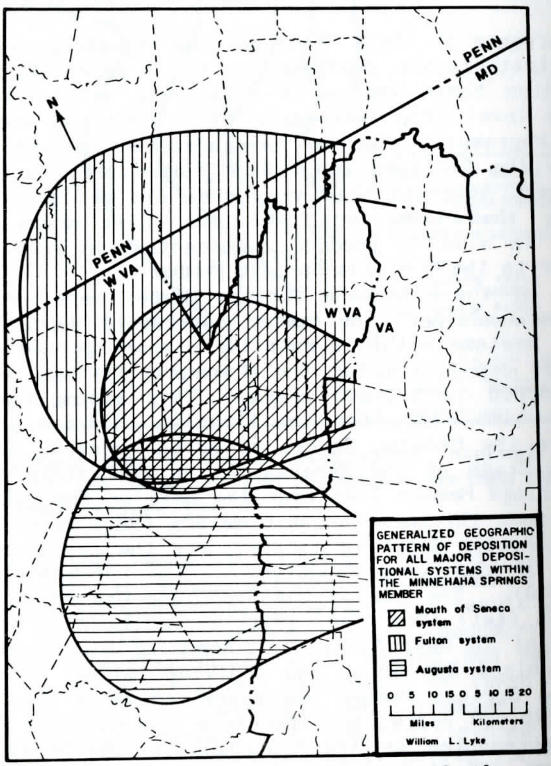

Figure 7. The generalized geographic location of the major depositional systems of the Minnehaha Springs Member. Modified slightly from Lyke, 1981 (Figure 16).

The Fulton and Augusta depositional systems were developed prior to the deposition of the Minnehaha Springs Member (Figure 6). The amount of time represented by the section between the base of the Minehaha Springs Member to the top of the Tully Limestone is assumed to be constant throughout the study area since the Tully (Hasson and Dennison, 1978) and the Minnehaha Springs Member (Lyke, 1981), are both considered to be brief time zones. The Fulton system was the most areally extensive and had the highest sediment accumulation rate of any depositional system in the field area during this time interval. As shown in Figure 6, the Mouth of Seneca system seems to have had little appreciable effect on the sediment pile in the geographic location of the present day Allegheny Front. The dominance of Fulton system deposition over Augusta system deposition could have resulted from the spatial relationship of a northeast trending outcrop belt to the approximate northward orientation of the paleo-shoreline, identified by Dennison and de Witt (1972) for Cohocton age rock units. This isopachous pattern could also be produced by the Fulton depositional system having a higher sedimentation rate in the study area than the Augusta system, perhaps due to an earlier development of the Fulton system.

The fan-shape geometry of the depositional systems and suprafans were identified by making isopach maps of the thickest bed within the Minnehaha Springs Member (Figure 5), and of the Minnehaha Springs Member (Figure 8).

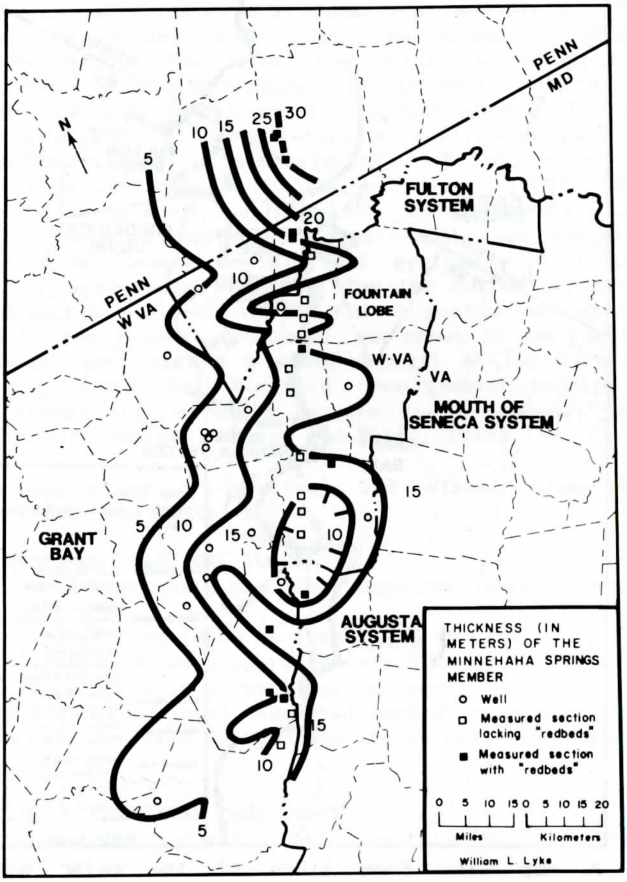

Figure 8. An isopachous map of the Minnehaha Springs Member based upon well data and measured sections from this study, Hasson and Dennison (1978), deWitt (1974), Martens (1945), Amsden (1954), and unpublished measured sections by John M. Dennison on file with the West Virginia Geological and Economic Survey at Morgantown, W. Va. Modified slightly from Lyke, 1981 (Figure 10).

A bilobed geometry for the Augusta system is shown in Figures 5 and 8. The Mouth of Seneca system contains a single suprafan lobe. The Fulton system may be represented by two suprafan lobes (Figure 8), with the dominant lobe centered at the Fountain (W-32) section. The Fountain suprafan may represent a separate depositional system related to the Frederick Lobe recognized by Jolley (1983). The Frederick Lobe is a depositional lobe within the Clearville Siltstone of the Middle Devonian Mahantango Formation. Perhaps deposition from the same stream system, through time, can be represented by the Frederick Lobe of Jolley (1983), the Fountain suprafan of the Minnehaha Springs Member, and the Route 50 Lobe (Kirchgessner, 1973). However, because accurate data to support these assumptions is lacking, the Fountain suprafan and all deposits north of the Elklick Run (W-27 section are considered to belong to the Fulton depositional system.

The geographic location of the three depositional systems observed in this study appear to have been relatively constant through much of the Upper Devonian. Figure 9 shows the form lines of these depositional systems

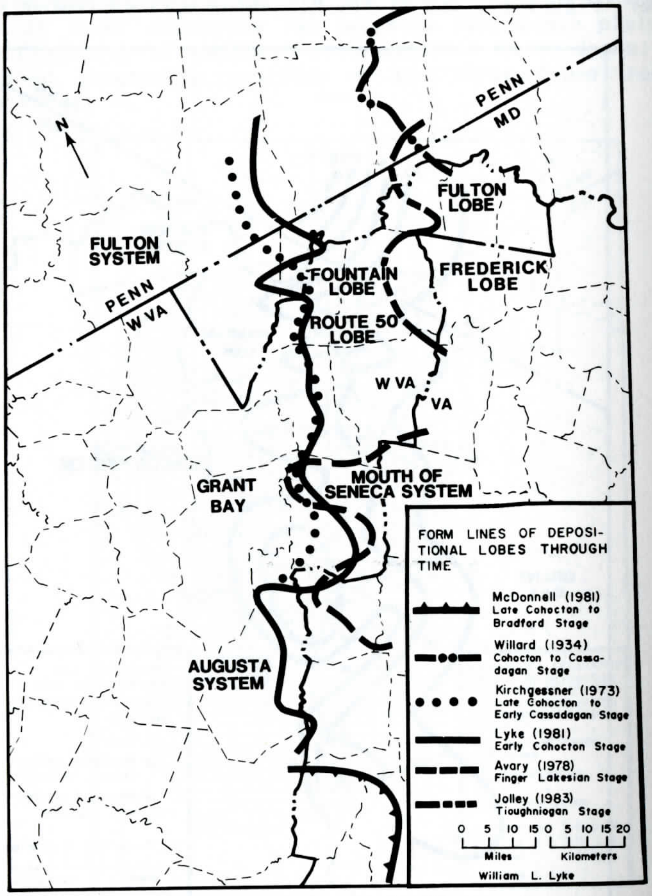

Figure 9. A map with form lines of the major depositional lobes demonstrating the geographic locations of each lobe through geologic time. Modified slightly from Lyke, 1981 (Figure 14).

through time as identified by six different studies in the Central Appalachians. Collectively, all of these studies recognized the same depositional systems. The Fulton system has existed at least from Middle Devonian Tioughniogan Stage (Jolley, 1983) to at least the top of the Upper Devonian Cohocton or early Cassadagan Stage Kirchgessner, 1973), approximately 10 million years (Harland *et al*, 1964). The Mouth of Seneca system appears to have been centered over the northwestern portion of Pendleton County, West Virginia from at least the Upper Devonian Finger

Lakesian Stage (Avary, 1978) to at least the Cassadagan Stage (Kirchgessner, 1973). The Augusta depositional system was recognized to exist in the Finger Lakesian Stage (Avary, 1978) and persisted at least to the Bradford Stage (McDonnell, 1981).

The progradation of the shoreline toward the west through time is also demonstrated by these six studies (Figure 9) along the Allegheny Front. Depositional environments range from lower mid-fan or outer fan in the Finger Lakesian Stage (Avary, 1978) to near-shore in the Cassadagan Stage (Kirchgessner, 1973) and shoreline in the Bradford Stage (McDonnell, 1981).

Submarine Fan Environment

Walker and Mutti (1973) assigned different turbidite facies to environments on modern slope-fan-basin plain systems. These turbidite facies and their characteristics are summarized in Table 1. The middle fan environment may be broken down into the channeled middle fan and depositional lobe middle fan area, the latter of which represents the dominant environment of the Minnehaha Springs Member. Characteristics of the depositional lobe environment are 1) the dominance of Facies D and C, 2) insignificant channeling, 3) more laterally continuous sand beds, and 4) (sand + silt)/shale ratios of approximately 1.0, which is less than the middle fan channeled environment (Walker and Mutti, 1973). The similarity of these characteristics with the Minnehaha Spring Members plus the absence of channeling, suggests a middle fan depositional lobe environment for the Augusta and Mouth of Seneca depositional systems. A lower middle fan environment is also generally suggested for the Fulton depositional system. However, some measured sections in the Fulton system represent environments farther upslope in the general fan model, as shown by amalgamated bedding at the Fountain (W-32) section and a slump deposit at the Turners Run (W-35) section in Mineral County, West Virginia. Amalgamated bedding may represent wide, shallow channel fill deposits because it is commonly associated with channeled mid-fan environments (Walker and Mutti, 1973).

Table 1. Characteristics of proximal and distal classical turbidites summarized from Walker and Mutti (1973).

FACIES C: PROXIMAL CLASSICAL TURBIDITES

1) Sandstone beds have sharp and flat bases regularly bedded with good lateral continuity.
2) Sandstones range from 10 cm to 1 meter in thickness.
3) (Sand + silt)/shale ratio approximately 5:1.
4) Amalgamation of beds is present but uncommon.
5) AE is the typical proximal classical turbidite sequence.
6) High proportion of beds in sequence begins with the A division at base.

FACIES D: DISTAL CLASSICAL TURBIDITES

1) Bases of sandstones and siltstones are sharp and flat, grading is prominent.
2) Grain size ranges from sandstones to siltstones 1 to 10 cm thick.
3) (Sand + silt)/shale ratio is 1:1 or less.
4) Biological activity preserved (tracks, trails, burrows).
5) Deposited on lower fan, basin plain, or as overbank deposits by proximal channel flows.
6) A and B divisions commonly are missing.

An outer fan environment is characterized by 1) no channels, 2) presence of thin, broad flows, 3) Facies D is dominant on active sedimentation portions of the fan, whereas pelagic and hemipelagic muds of Facies G occur on inactive portions, and 4) (sand + silt)/shale ratios are less than 1.0

on inactive portions, and 4) (sand + silt)/shale ratios are less than 1.0 (Walker and Mutti, 1973). Sections with less than 50% coarser-grained rocks exist at Route 250 (V-2), Maysville (W-25), Elklick Run (W-27), and Madley (P-1). These four sections more closely approximate outer fan deposits than any other measured sections in this report.

The shaley Facies G deposits recognized above and below the Minnehaha Springs Member, as well as between the two coarser-grained packets present in the Zeke Run (W-17), Brushy Run (W-16), Ketterman Knob (W-15), and Briery Gap Run (W-14) sections, represent basin plain or inactive fan deposits.

FAUNA AND COMMUNITY STRUCTURE

Modeling of the community structure associated with the Minnehaha Spring Members was attempted using non-parametric Spearman Correlation Coefficients with a 95% level or greater of significance. The computer program may be found in the Statistical Package for the Social Sciences (SPSS) at Triangle Universities Computation Center (TUCC) at the University of North Carolina, Chapel Hill. Only those samples containing more than one species were used in the program, because single species occurrences would not affect the correlations. Samples containing single species were collected from silty shale lithologies and included *Douvillina* sp., *Dalmenella* sp., *Pseudoviculopectin striatus*, and *Sandbergeroceras chemungensis*. By far the most abundant members of the Minnehaha Springs taxonomy were *Ambocoelia umbonata* and pelmatozoan fragments.

Three separate communities were recognized (Figure 10): the Burrowing Community (Community I), the Pelmatozoan-*Ambocoelia*-*Chonetes* Community (Community II), and the *Loxonema* Community (Community III). The Burrowing Community, containing non-branching and branching horizontal burrows and burrow holes, was prevalent in silty shales, shales, and very thinly bedded siltstones and sandstones within the Minnehaha Springs Member. Trails, such as *Pteridichnites biseriatus*, were also conspicuous in the silty shales. *Pteridichnites* is thought to be the crawling trail of an arthropod or annelid (Moore, 1962, p. W210). Of the three communities present, only Community I is associated with shaly lithologies. Therefore, the Burrowing Community represents the typical basin plain or outer fan community.

Closely spaced episodic events which deposited the Minnehaha Springs Member siltstones and sandstones are responsible for the other two communities. The Pelmatozoan-*Ambocoelia*-*Chonetes* Community is located in the Grant Bay area from the Fountain (W-32) to Route 250 (V-2) sections. The highest faunal densities for this community occur at Maysville (W-25) and Route 250 (V-2). These two locations are on the edge of Grant Bay between the Fulton and Augusta systems. The *Ambocoelia* and Pelmatozoan fragments tend to be located in high concentration zones approximately 3 centimeters thick both at the base and at the top of coarse-grained siltstone beds. Abrasion of the *Ambocoelia* shells indicates minor shell transport. the Pelmatozoan-*Ambocoelia*-*Chonetes* Community, particularly when present at the top of a coarser-grained bed, may represent an opportunistic community. *Ambocoelia* has been identified by Diehl (1980) as an opportunistic species and therefore would colonize an area after the original community was destroyed by a turbidity flow. The original community in this instance was the Burrowing Community.

The *Loxonema* Community represents a more typical upslope fauna which has been swept down-slope by turbidity flows. It does not represent a natural community, but rather a thanatocoenosis. The community, which is found locally at the Maysville (W-25) section, was collected from the base of a coarse-grained siltstone bed. Shell abrasion was not apparent. Community III is more taxonomically diverse than the other two communities, and contains six different species.

These three communities can be related to the progradation of lower mid-fan environments over the outer fan or basin plain environments (Figure 10).

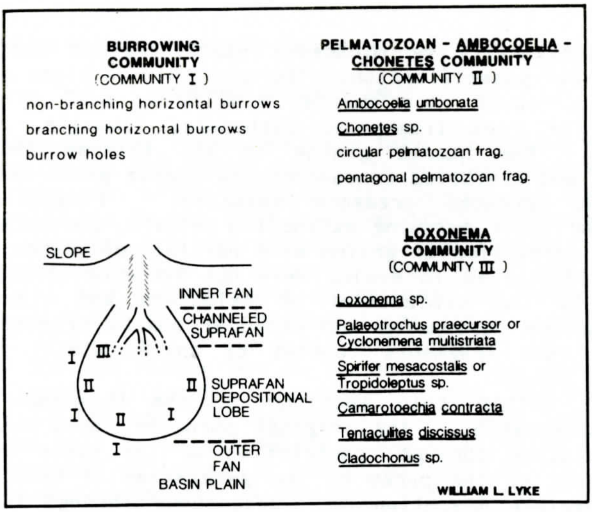

Figure 10. The relationship of the Minnehaha Springs Member faunal communities to generalized submarine fan depositional environments.

of low sedimentation rates in an outer fan or basin plain environment. During episodic turbidity flows, individuals of the *Loxonema* community were localized near the discharge of channels on the suprafan as flow velocity diminished. Immediately after rapid sedimentation ceased, the opportunistic Pelmatozoan-*Ambocoelia-Chonetes* Community colonized local areas on top of coarser-grained beds. With the return of lower energy conditions, the Burrowing Community once again became dominant. Therefore, Community I can be found in both the suprafan and outer fan environments (Figure 10).

Seventeen species were found in the Fulton system deposits, whereas the Augusta system contained 10 species. Grant Bay, including the Mouth of Seneca system, contained 14 species. Not only does the species diversity increase towards the northeast, but the only pelecypods observed occurred at LaVale (M-2) in the Fulton system. A greater pelecypod to brachiopod ratio (Bretsky, 1969) and an increase in species diversity in offshore environments may indicate a more shoreward position of the locality. This corresponds well with the more nearer-shore depositional environments interpreted earlier to exist on the Fulton depositional lobe.

PETROLOGY

A total of nine thin-sections from siltstones and sandstones were analyzed to provide a general petrographic description of the Minnehaha Springs Member. Each of the nine thin-sections were located randomly within the Minnehaha Springs Member and were chosen from nine separate measured sections so that each of the three major depositional systems were represented by three thin-sections. One hundred point counts were made on each thin-section. This petrographic study was not designed to be of statistical significance, but rather to provide a general petrographic description of the Minnehana Springs Member of the Scherr Formation. A point count determination using 100 counts has a 95 percent confidence reliability of +/-10.0% for grains with 50% abundance, a reliability of +/-8.0% for grains with 75 or 25 percent abundance, a reliability of +/-6.0% for grains with 90 or 10 percent abundance, and a reliability of +/-5.0% for grains of 95 or 5 percent abundance, according to calculations to the nearest whole percentage made from binomial sampling theory (Dixon and Massey, 1969, pp. 245-250, 502-504).

The coarser-grained beds of the Minnehaha Springs Member can be classified texturally as micaceous coarse-grained siltstone according to Picard's (1971) classification. The nine thin-sections contain an average of 62.5% quartz, 16.6% clay matrix, 8.0% mica, 4.2% feldspars, and smaller percentages of rock fragments, carbonate, limonite, and opaque minerals (Table 2). The quartz grains in all thin-sections were angular to subangular, and exhibited concavo-convex quartz grain sutures, indicating an event which produced pressure solution. Plagioclase feldspars were identified by their twinning extinction pattern, and potassium feldspars were tentatively identified as grains with partial replacement by carbonates. The carbonates identified in grains were not syndepositional but appear to have originated by the replacement of feldspars and clay matrix. Fracture fillings include carbonates and euhedral quartz crystals. Sedimentary and metamorphic rock fragments consist of shale, slate, phyllite, and a few fragments of schist.

Porosity within thin sections of the Minnehaha Springs Member is visually non-existent. Any original porosity would have been destroyed by pressure solution through constriction of the pores by compressive forces and/or filling of the pores by the generation of secondary matrix (Kuenen, 1966). Possible migration of hydrocarbons through the Minnehaha Springs Member may have occurred before the pressure solution event, during the early stages of pressure solution, or after the pressure solution occurred through fractures propped open by euhedral quartz crystals such as those observed at the Route 250 (V-2) sections. However, carbonaceous residues were not observed in any thin-sections.

The compressive forces needed for pressure solution could have been supplied by deep burial of the sediment after deposition which is a common diagenetic phenomena, and/or by folding (Geiser, 1974; Groshong, 1975),

Table 2: Summary of petrographic data for each major depositional system within the Minnehaha Springs Member.

	Depositional System SW		NE	
	Augusta	Mouth of Seneca	Fulton	Average
% quartz	54.3	64.3	69.0	62.5
% feldspar	1.7	4.7	6.3	4.2
% mica	13.0	5.3	5.7	8.0
% rock fragments	3.3	1.3	4.0	2.9
% clay matrix	19.0	18.0	12.7	16.6
% limonite	4.3	2.7	1.3	2.8
% opaque minerals	4.3	2.3	1.0	2.5
% total carbonate	0.7	4.0	5.7	3.5

Each number is the arithmetic mean percentage from the mineral category from three thin sections (100 point counts from each) of the three major depositional systems.

perhaps by the Allegheny Orogeny (Dean and Kulander, 1978).

A summary of the petrographic data for each major depositional system is presented in Table 2. Table 2 shows a general decrease in the averaged percentage values for quartz, feldspar, and replacement calcite towards the southwest, whereas the averaged percentages for micas, clay matrix, and opaque minerals increased. These geographic trends in mineral content may represent 1) an increasingly lower rank metamorphic source area toward the southwest, such as that identified by Kirchgessner (1973) for the upper part of the Cohocton Stage in the Foreknobs Formation of the Greenland Gap Group along the Allegheny Front, or 2) the angular relationship between the present-day northeast trending outcrop belt and the northward trending Cohocton age paleoshoreline recognized by Dennison and de Witt (1972).

ECONOMIC SIGNIFICANCE

The Brallier and Chemung Formations have been identified (Glaeser, 1979) as hydrocarbon reservoirs and source beds. In ten north-central and western counties of West Virginia, wells reaching into Chemung and Brallier coarser-grained bundles below the Benson sand have dominated exploratory drilling in the past decade. Present drilling activity in West Virginia is focusing on a series of siltstones in the north-central portion of the state, located stratigraphically in the marine Devonian between the Benson and Sycamore zones. These units have various informal local names, usually designating the discovery well or the name of the driller (Patchen *et al*, 1980).

The Minnehaha Springs Member may be the nearer-shore correlative of the gas producing Sycamore Siltstone identified in the subsurface of north-central West Virginia by Nock and Patchen (1975). Nock and Patchen (1975) report the Sycamore grit to be 427 meters (1400 feet) below the Benson sand and the lowest mappable siltstone or sandstone unit in the Upper Devonian sequence of interbedded sandstones, siltstones, and shales.

Hydrocarbon production from other mid-fan turbidity flows, such as the Benson sand (Cheema *et al*, 1977), is encouraging for the economic potential of the Minnehaha Springs Member. The best probable location for hydrocarbon exploration in the Minnehaha Springs Member is in the west-central and north-western portion of West Virginia and south-central Pennsylvania for the following reasons: 1) There is an apparent coarsening of the unit westward in the Fulton system as the Minnehaha Springs Member contains 70% sandstone in the J. F. and Harrison Sisler No. 1 (W-23) well (Figure 2) but only 50-55% siltstone in outcrop to the east. The coarser nature of the member in the well may suggest the presence of a channel. A lithogic description of the well is presented in Amsden (1954). 2) There is an apparent overlapping of deposits from the Mouth of Seneca and Augusta systems, and perhaps the Fulton system, in the Randolph 102 (W-9) and Randolph 58 (W-11) wells. The presence of intervening shales as seal rock would enhance the possibility of a hydrocarbon reservoir.

Berger (1978) suggested that the Glady, Horton, Blackwater, and Deer Park anticlines are favorable drilling sites for the Brallier Formation. The Minnehaha Springs Member should be a favorable drilling target on these anticlines as well.

ACKNOWLEDGEMENTS

This paper is the result of the author's M.S. thesis at the University of North Carolina, Chapel Hill. Research was partially funded by the Martin Fellowship (U.N.C.-C.H.), West Virginia Geological and Economic Survey, the National Association of Sigma Xi, the Smith Fund (U.N.C.-C.H.), and Columbia Gas Transmission Corporation for thin section preparation.

Many thanks are due John M. Dennison for suggesting this topic for study, for his expert advice during the project, and for his critique of this manuscript prior to publication. Richard M. Jolley participated in many relevant discussions and, along with Frederick P. Lyke, contributed valuable field assistance. Wallace de Witt, Jr., provided the author with outcrop localities in south-central Pennsylvania. Mr. de Witt and Katharine Lee Avary reviewed this manuscript and submitted very helpful and pertinent suggestions. The contribution of drafting expertise for this manusript by Chevron U.S.A., Inc. is greatly appreciated.

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STRATIGRAPHY, DEPOSITIONAL ENVIRONMENTS, AND REGIONAL DOLOMITIZATION OF THE BRASSFIELD FORMATION (LLANDOVERIAN) IN EAST-CENTRAL KENTUCKY: DISCUSSION AND REPLY

DISCUSSION

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The Brassfield Formation has been called "one of the remarkable and distinctive sequences of all Paleozoic time" (Ham and Wilson, 1967, p. 354). Gordon and Ettensohn (1984) have made an interesting and useful contribution to our understanding of this widespread and distinctive stratigraphic marker, particularly in their detailed observations of sedimentary features within the Brassfield. Two areas of their analysis, however, call for critical comment: (1) the age and definition of the Brassfield Formation and (2) the interpretation of the paleogeography of the unit during its formation, which has implications for their dolomitization model.

Gordon and Ettensohn (1984, p. 101, 103) characterize the Silurian exposures in eastern Kentucky as a sequence of interbedded dolostones and shales of Early Silurian (Llandoverian) age which are divisible into two formations, the Brassfield Formation of Early and Middle Llandoverian age, and the overlying Crab Orchard Formation. Detailed mapping of these rocks during the recent Kentucky Geological Survey - U.S. Geological Survey cooperative mapping program (1960-1978), summarized by McDowell (1983), suggests a somewhat different stratigraphic framework.

The Silurian sequence described by Gordon and Ettensohn (1984) is of both Early and Middle (Albion-Niagaran) Silurian age, and the Brassfield Formation is of Early Llandoverian age, as indicated by conodont studies by various workers (see McDowell, 1983, p. 19-23 and figure 16). Locally this sequence is overlain by a third Silurian formation, the Bisher Dolomite (Wenlockian).

The traditional two-fold subdivision of the pre-Bisher Silurian into the Brassfield and Crab Orchard Formations has proven to be unworkable north of Bath County, Kentucky, where the top of the Brassfield was defined by a faunal horizon, the "bead bed" of Rexroad and others (1965), which is not practical for mapping or subsurface studies. This problem is described and illustrated by Gordon and Ettensohn (1983) themselves: Their "upper massive beds," at the top of the Brassfield, pinch out into shales (p. 109), and on their figure 3 and 4 no lithologic break is shown between the two formations in columns seven and eight. Moreover, the "upper massive sandstone" of column six is actually the Oldham Member of the "Crab Orchard Formation" (McDowell, 1983, plate 1, Bluebank section). Because the Brassfield has no lithologically defined top north of Bath County and is therefore not mappable there, and is relatively thin to the south, McDowell (1983) established a new two-fold subdivision of the Brassfield-Crab Orchard interval which reflects the actual map units used during the quadrangle mapping program. These units were named Drowning Creek Formation (predominatly dolomite) below and Alger and Estill Shale above (McDowell, 1983, p. 7-19) (see figure 1). The Brassfield Formation in eastern Kentucky was reduced in rank to member of the Drowning Creek Formation. The specific nomenclature used in this area can be debated, but recognition of the problem of faunal definition of the top of the Brassfield (the "bead bed") is essential. This problem has been discussed more fully elsewhere (O'Donnell, 1967, p. 34-35; McDowell, 1983, p. 7-8).

A second difficulty is the proposed paleogeography of the lower Silurian rocks. Gordon and Ettensohn (1984, p. 107, 110, and figure 6) describe the Brassfield as a near-shore lithofacies formed on the flank of a "proto-Cincinnati Arch" which constituted a land area during Brassfield time. The

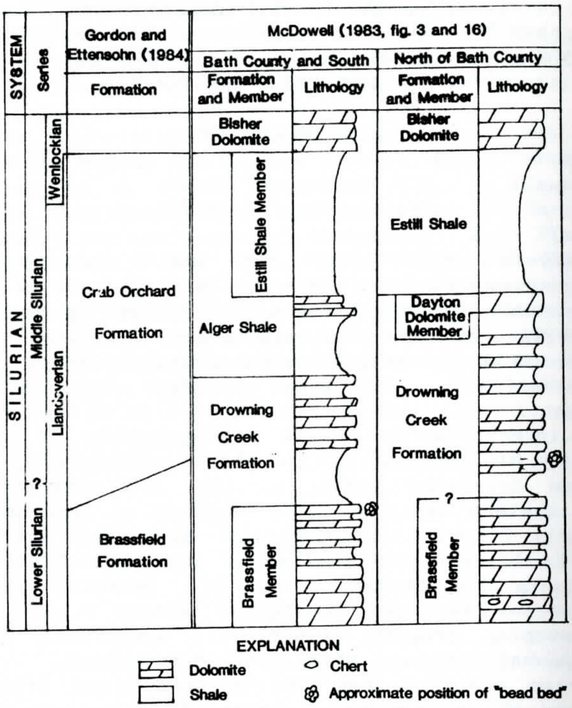

Figure 1. Generalized columnar sections of Silurian rocks exposed east of the Cincinnati Arch in Kentucky.

lithology and distribution of the Brassfield in Kentucky suggest that this is very unlikely for several reasons. (1) Coarse clastic sediments are present, and increase in amount, eastward in the subsurface of eastern Kentucky and West Virginia (Perry, 1962; Horvath, 1967); there is no increase in clastic sediments westward towards the arch. (2) The basal Silurian does not overstep progressively older Ordovician units towards the axis of the arch, as would be expected if it were transgressing across an eroded tectonic uplift (Weir and others, 1984). (3) The Brassfield is widely distributed on the west flank of the arch (Peterson, 1981), and several isolated inliers of Brassfield occur near the axis of the arch in central Kentucky, such as at Scrubgrass Creek, in Boyle County, and on Jeptha Knob, in Shelby County (McDowell and Peterson, 1980; McDowell, 1983; Cressman, 1981). It seems likely, therefore, that the Brassfield originally extended across the arch, which apparently only began to form in Early Silurian time (McDowell and Peterson, 1980; McDowell, 1983, p. 24) and probably had no effect on sedimentation until some time after Brassfield deposition (Freeman, 1951, p. 9; O'Donnell, 1967, p. 116-118). This suggests that no land area can be associated with the arch in central Kentucky during Brassfield time as postulated by Gordon and Ettensohn (1984, p. 110, figure 6). Therefore an offshore sand bar belt probably was not present in the manner they propose. Moreover, the transition from dolomite to limestone takes place from south to north, across the arch, as noted by Gordon and Ettensohn (1984, p. 101-102) themselves, and not parallel to it. Thus, whatever the merits of the model otherwise, Early Silurian paleogeography does not appear to support the mixing-zone dolomitization mechanism suggested by Gordon and Ettensohn (1984, p. 110-111), and this dolomitization does not seem clearly related to the present Cincinnati Arch.

REPLY TO DISCUSSION

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McDowell has suggested two areas in our analysis of depositional environments in the Brassfield Formation (Gorden and Ettensohn, 1984), which require further explanation, and we are glad to provide this.

First, McDowell calls to question our age and definition of the Brassfield Formation. In his discussion, he states that we "characterize the Silurian exposures in eastern Kentucky as a sequence of interbedded dolostones and shales of Early Silurian (Llandoverian) age which are divisible into two formations, the Brassfield Formation of Early and Middle Llandoverian age, and the overlying Crab Orchard Formation." What we really say (Gorden and Ettensohn, 1984) is that "Lower Silurian rocks in east-central Kentucky can be subdivided into two formations and several widely correlatable members (bottom p. 103)," and that the Brassfield is Early-Middle Llandoverian in age (p. 101). Neither of these statements is incorrect based on the information available at the time of our writing. Clearly, in the older nomenclatorial scheme used in our paper, Lower Silurian rocks encompass the Brassfield Formation and lower portions of the Crab Orchard Formation; upper parts of the Crab Orchard are Middle Silurian (Niagaran). Despite McDowell's assertions to the contrary, our two interpretations of this point do not differ. Perhaps McDowell was confused by an error in our Figure 2, in which a horizontal line delimiting the upward extent of Lower Silurian rocks was omitted in the far right-hand column.

Our definition of the Brassfield as Early-Middle Llandoverian in age is based on the work of Berry and Boucot (1970); we did not have access to McDowell's (1983) data. Nonetheless, as late as 1984, Rexroad and Kleffner indicated that Early Llandoverian ages are still probable for parts of the Brassfield, although most of the unit is probably Middle Llandoverian and perhaps younger.

As for naming the Silurian rocks, we used the traditional Brassfield-Crab Orchard nomenclature, because at the time our paper was submitted and reviewed, McDowell's nomenclature was not yet available. Had it been available, however, we doubt that we would have used it. As Figure 1 in McDowell's preceding discussion shows, the Drowning Creek Formation is not a homotaxial unit. In evnironmental studies like ours where it is convenient to refer to a given body of rock with a single name, the use of a unit name, which does not refer to the same body of rock throughout its distribution, only compounds problems. Based on McDowell's (1983) discussion of age and correlation in the Drowning Creek (p. 19), we suspect that the name changes were sugggested by problems in defining the top and age of the Brassfield. For example, McDowell (1983) stated, "on the basis of conodont data, the Brassfield Member of the Drowning Creek of eastern Kentucky is restricted to the lower Llandoverian, and strata of middle Llandoverian age, placed in the Brassfield by Berry and Boucot (1970), are here regarded as those constituting the upper part of the Drowning Creek." The paragraph ends with the statement that "The Brassfield-Drowning Creek interval may thus be of about the same age throughout the Cincinnati Arch area."

We do not believe that nomenclatorial changes should have any basis in age. Although the diachronous nature of the Brassfield has been known for nearly 20 years (Rexroad, 1967; Berry and Boucot, 1970), the Brassfield and overlying Crab Orchard are well established and much used units; they are easily distinguished throughout most of their distribution. As McDowell pointed out, however, problems arise in northeastern Kentucky (north of Fleming County, we believe) where upper parts of the Brassfield pinch out into the Crab Orchard. At this point, we followed the usage of Rexroad and others (1965) and chose the "bead bed" as the top of the Brassfield. Inasmuch as this is the use of biostratigraphy to define lithostratigraphic units, it is inappropriate. However, we also believe that solving the Brassfield-Crab Orchard boundary problem in a local area by combining well established homotaxial formations into a unit, that is not homotaxial across its distribution, probably is equally inappropriate and certainly unnecessary. To complicate matters more, nomenclature of Lower and Middle Silurian rocks (Brassfield, etc.) on other parts of the same outcrop belt in west-central Kentucky and southwestern Ohio will remain unchanged. We suggest that a more acceptable solution to the boundary problem is merely the formal redefinition of the top of the Brassfield as the top of the last major dolostone.

Secondly, we believe that McDowell's comments on our paleogeography show that he did not understand what we tried to convey. Apparently we were unclear. We were aware of the three lines of evidence he cites, but it is clear that he does not fully explore all the implications of each line: (1) We certainly agree that subsurface equivalents of the Brassfield coarsen eastward toward the orogen, and that there is no increase in clastics toward the arch. However, even if the arch had been completely emergent, few clastics would be expected in the Brassfield, because no clastic source existed on the arch. The Ordovician sediments below the Brassfield are all carbonates and calcareous shales. At the most, such a carbonate terrane would have produced a few carbonate lithoclasts, which are found locally in the basal Brassfield. Even though the amount of clastics does not increase toward the arch, what McDowell fails to understand from our paper is that carbonate grain size in the Brassfield (prior to dolomitization) does increase in that direction. We will come back to this later. (2) We agree that the Brassfield does not overstep progressively older units (except the Belfast Member) toward the axis of the arch. This suggests to us that though the arch was uplifted, uplift was mild and episodes of subaerial exposure were short. In fact, the presence of glauconite and phosphatized clasts in the base of the Brassfield suggests subaequeous erosion and little sedimentation during the Ordovician-Silurian transition. We argue, therefore, that the "proto-Cincinnati Arch" was uplifted but seldom emergent. (3) We are well aware that the Brassfield crops out on the western side of the arch, and we likewise believe that the Brassfield was deposited across the arch; after all, it is present in outliers on the axis of the arch (Foerste, 1931). Nowhere in the paper do we suggest or infer that the arch was exposed throughout Brassfield deposition; this was wholly McDowell's inference from our paper. Although we do not specifically deal with this problem, our Figure 6 suggests that during early parts of the Brassfield transgression, parts of the proto-Cincinnati Arch were covered with tidal-flat and lagoonal sediments like those found in the Belfast Member. Periodic exposure may be expected in such environments. More importantly , however, few transgressions are geologically "instantaneous" or continuous. It would have taken time for Brassfield deposition to migrate across the arch, and as we demonstrated in the paper through the presence of small disconformities and repeated transgressive sequences, the transgression was not continuous, but interrupted by smaller regressive events. Either of these situations may have resulted in brief periods of exposure. The only instance where we do specifically call on exposure for dolomitization is in the sand belt-shoal environment of our transgressive continuum. We do not think this is unreasonable, as sands in such environments may accrete so fast during storms as to build up above sea level for short periods of time (Ball, 1967). One can argue that none of these situations provided enough exposure during the Early-Middle Silurian to effect mixing-zone dolomitzation. This may be the case, but as we argued in our paper, the Early-Middle Devonian may have been even more opportune time for mixing-zone dolomitization, as the greater duration of uplift would have made exposure even more probable.

Finally, we dispute McDowell's statement that the Cincinnati Arch or a proto-arch had no effect on sedimentation until after Brassfield depositon. We provide three lines of evidence that counter McDowell's assertion: (1) the Brassfield thins southwestwardly as the outcrop belt approaches the axis of the arch; (2) the Brassfield becomes coarser-grained in the same direction reflecting higher energies in shallower water near the arch; and (3) younger parts of the Brassfield overstep the Belfast Member southwestwardly toward the arch. We suspect that the Belfast also was deposited on and near the arch, but that higher-energy Brassfield environments eroded it away during transgression across the arch.

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